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MINERALISATION NEAR THE NORTHEAST MARGIN OF THE NELSON BATHOLITH, SOUTHEAST BRITISH COLUMBIA.

by

© SIMON BRAME

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES

AND RESEARCH IN PARTIAL FULFILMENT OF THE

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MASTER OF SCIENCE

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FALL , 1979

THE UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled "Mineralisation near the northeast margin of the Nelson Batholith, southeast British Columbia", submitted by Simon Brame in partial fulfilment of the requirements for the degree of Master of Science.

R.S. Lambert

Supervisor

Date 7 May 79

TO "ALDA BABES"

ABSTRACT

The Scranton, Flint and Vigilant lead-zinc deposits near the northeast margin of the Nelson Batholith have been investigated and related to the better documented Bluebell deposit. The study confirms the "classic" idea of mineralisation being related to the cooling stages of the batholith and therefore having an age of Middle Jurassic.

The sulphur isotope results (30 analyses) from this study suggest that the ore sulphur is igneous in origin with a small sedimentary component. The isotopic values at the Vigilant deposit range from -7.6‰ to -8.8‰ for sphalerite and -9.5‰ to -11.2‰ for galena and at the Scranton deposit from -2.8‰ to -5.5‰ for sphalerite and -2.8‰ to -6.3‰ for galena. These values, showing $\delta^{34}S$ depletion relative to a meteoritic standard, contrast strongly to sulphur isotopic values at the Bluebell deposit. The temperatures obtained from galena-sphalerite pairs are; Vigilant (160 - 390°C), Scranton (300 - 510°C) and Bluebell (650°C). The Flint deposit samples show apparent re-equilibration at lower temperatures. Each deposit has unique sulphur isotope characteristics which indicate that the mineralising

solutions were well mixed prior to deposition but the source areas were localised and at least partially separated.

A study of the fluid inclusions from sphalerite and quartz shows that the mineralising solutions were relatively hot, indicative of a primary hydrothermal origin. The fluid inclusion temperatures from Scranton (380 - 465°C) and Vigilant (370-390°C) correspond well with the sulphur isotope temperature determinations using galena-sphalerite pairs. The presence of daughter crystals indicates that the solutions were made up of complex saline brines with up to about 40 NaCl equivalent weight%. Heterogeneous trapping ("boiling") of gas-rich and brine-rich fluid inclusions at Flint indicates that the deposits were precipitated close to the surface (500m?). The solutions were probably injected into the fissures at high velocities, depending on the varying pressure conditions produced by simultaneous tectonic movement. Secondary fluid inclusions from Scranton show a temperature of deposition similar to that of the primaries, again indicating tectonic movement during deposition. The ores at Scranton and Flint, within the Nelson Batholith, show very distinct cataclastic and replacement textures.

7 whole rock X.R.F. analyses on various phases of the Nelson Batholith confirm that the plutonic complex is distinctly calc-alkaline in character.

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My presence at the University of Alberta is due to the efforts of Dr. Richard Lambert who, as supervisor, also advised the area of study and assisted me throughout the thesis for all of which I am sincerely grateful.

Dr. Roger Morton provided the geological framework to undertake a thesis in economic geology and also the motivation to complete it.

My thanks are also due to Mr. H. Hemmerich, of HEM Mines, who allowed me to visit Scranton Mine and showed me around the underground workings. The use of the scanning electron microscope in the Entomology Department and the Nicolet absorption spectrometer in the Chemistry Department is gratefully acknowledged. Help was kindly provided by Dr. H. Baadsgaard and Alex Stelmach with the atomic absorption determinations, Dr. D.G.W. Smith and Steve Launspach with the microprobe analyses, and Frank Dimitrov with the photographic plates and drafting. The X.R.F. analyses were performed at Durham University, England, by Dr. J. G. Holland and Ron Hardy to whom I am indebted. Paul "Willie" Wagner

proved to be an extremely willing and very vocal field assistant. Without the financial support provided by the Department of Geology the thesis could not have been undertaken. Without Dianne Storheim's excellent typing I would still be in the Allan Library "whiting out." Jenni Blaxley helped with the specimen preparation and provided "light-relief." Dr. Roy Krouse kindly made available his mass spectrometer and his expertise.

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TABLE OF CONTENTS

CHAPTER		PAGE
I	GENERAL GEOLOGY OF THE KOOTENAY ARC	1
II	GEOLOGY OF THE STUDY AREA	5
	a) The Nelson Batholith	5
	b) The Sedimentary Succession	21
	i) Badshot Formation	21
	ii) Lardeau Group	21
	iii) Milford Group	22
	iv) Kaslo Group	23
	v) Slocan Group	23
III	LEAD-ZINC DEPOSITS OF THE KOOTENAY ARC	26
	a) Concordant	27
	b) Transgressive	27
	i) Bluebell	28
	ii) Ainsworth	30
	iii) Slocan	31
IV	HISTORY OF EXPLORATION AND DEVELOPMENT AT THE SCRANTON MINE	33
V	GEOLOGICAL DESCRIPTION OF THE SCRANTON DEPOSIT	41
VI	MINERALOGY AND PARAGENESIS AT THE SCRANTON DEPOSIT	54
	Associated Deposits	
	i) Flint	63
	ii) Vigilant	64

CHAPTER		PAGE
VII	ELEMENTAL GEOCHEMISTRY OF THE SULPHIDES...	67
	a) Scranton.....	67
	b) Flint.....	69
	c) Vigilant.....	70
VIII	SULPHUR ISOTOPE STUDIES.....	72
	Factors determining sulphur isotope fractionation	
	a) Temperature.....	73
	b) Biologic Activity.....	74
	c) Solution Chemistry.....	76
	Previous work done on Sulphur Isotopes....	77
	Present work.....	78
	Temperature.....	83
	Conclusions from the Sulphur Isotope Analyses.....	85
IX	FLUID INCLUSION STUDIES.....	89
	Description of the Fluid Inclusions....	89
	Results of Crushing Experiments.....	98
	Determination of Organics.....	99
	Salinity Estimates.....	100
	Heating Experiments.....	101
	Depth of Deposition at Flint.....	103
	Temperature Determinations.....	104
	Conclusions.....	107

CHAPTER PAGE

X ORIGIN OF THE MINERAL DEPOSITS..... 110

***

BIBLIOGRAPHY..... 120

APPENDIX 1. X-RAY FLUORESCENCE ANALYSES..... 130

i) Major Elements (%)..... 130

ii) Minor Elements (ppm)..... 131

iii) Key to Specimen numbers for X.R.F. Analyses..... 132

APPENDIX 2. ATOMIC ABSORPTION..... 133

i) Principles involved..... 133

ii) Analytical Procedure..... 133

a) Sulphides..... 133

b) Granites, and individual K-feldspars..... 134

iii) Tabulated Results.....

a) Sulphides..... 135

b) Granites and K-feldspars..... 136

iv) Correlation with U.S.G.S. Standards..... 137

APPENDIX 3. SULPHUR ISOTOPES..... 138

i) Preparation of Samples..... 138

ii) Collection of SO_2 gases..... 140

iii) Mass Spectrometer measurements.. 141

iv) Sample computer printout..... 142

v) Calibration with standards..... 142

CHAPTER PAGE

APPENDIX 4. FLUID INCLUSIONS.....

i) Specimens Used..... 144

ii) Sample preparation..... 144

iii) Microthermometry method..... 144

iv) Crushing Experiments..... 145

v) Volumetric Salinity Determinations 146

LIST OF TABLES

Table	Description	Page
1	Sample Assay Returns for ore from the Scranton Mine	40
2	Minor Element Geochemistry of Sphalerite	68
3	Minor Element Geochemistry of Pyrrhotite	71
4	Variation in $\delta^{34}S$ values	80
5	Temperature Calculations from Galena-Sphalerite Pairs	84

LIST OF FIGURES

Figure	Description	Page
1.	Geological map of the southern part of the Kootenay Arc showing the location of the field area.	2
2.	Geological map of the study area.	6
3.	AFM plot of plutonic rocks from the northern margin of the Nelson Batholith.	16
4.	MgO/SiO_2 Larsen diagram of plutonic rocks from the northern margin of the Nelson Batholith.	17
5.	TiO_2/FeO plot of plutonic rocks from the northern margin of the Nelson Batholith.	18
6.	Property and workings, Scranton Mine.	34
7.	Graph showing the results of calculations and experiments to determine the variation with temperature of sulphur isotope fractionation between galena and sphalerite.	75
8.	Plot of the sulphur isotope value determinations for galena and sphalerite.	79
9.	Plot of δ ‰ versus $\Delta\delta^{34}S$ for the sulphide pairs.	81
10.	Temperature correction for a 25 percent or greater NaCl solution as a function of homogenisation temperature and pressure.	106

LIST OF PHOTOGRAPHIC PLATES

Plate	Description	Page
I-1	View of Keen Creek looking Southeast	8
I-2	Contact between the Nelson Batholith and the Slocan Sediments	8
II	The Nelson Batholith	9
III	Scranton Ore	49
IV	Cataclastic and Replacement Textures	51
V	Vigilant Ore	53
VI	Electrum within Scranton Ore	57
VII	Argentite within Scranton Ore	59
VIII	Fluid Inclusions from Flint	93
IX	Fluid Inclusions from Scranton and Vigilant	95
X	Scanning Electron Microscopy	97

CHAPTER I

GENERAL GEOLOGY OF THE KOOTENAY ARC

The Kootenay Arc is a curving belt of complexly deformed sedimentary, volcanic and metamorphic rocks, over 250 miles long, from north of Revelstoke south into northeast Washington. The essentially conformable succession, which generally youngs westward, ranges from earliest Cambrian to Jurassic in age. To the east the succession passes downwards into older rocks of the Purcell Anticlinorium and to the west is bounded by the allochthonous Shuswap Metamorphic Complex. Granite plutons and associated intrusives (eg.the Jurassic Nelson Batholith) cut the older deformed stratigraphic succession. The most significant stratigraphic marker is the Lower Cambrian Badshot Limestone (in the Kootenay Lake area) and its southern equivalent, the Reeves Limestone, which occurs south of Nelson in the Salmo district. The extent of this marker horizon can be seen in Figure 1. Deformation within the Kootenay Arc is polyphase with the existence of at least three periods of folding. The structures are extremely complex in detail with overturned, strongly asymmetrical, isoclinal folds with axes parallel to the local trend of the arc. Field work has shown that the different phases of folding actually belong to one continuous process extending over a long

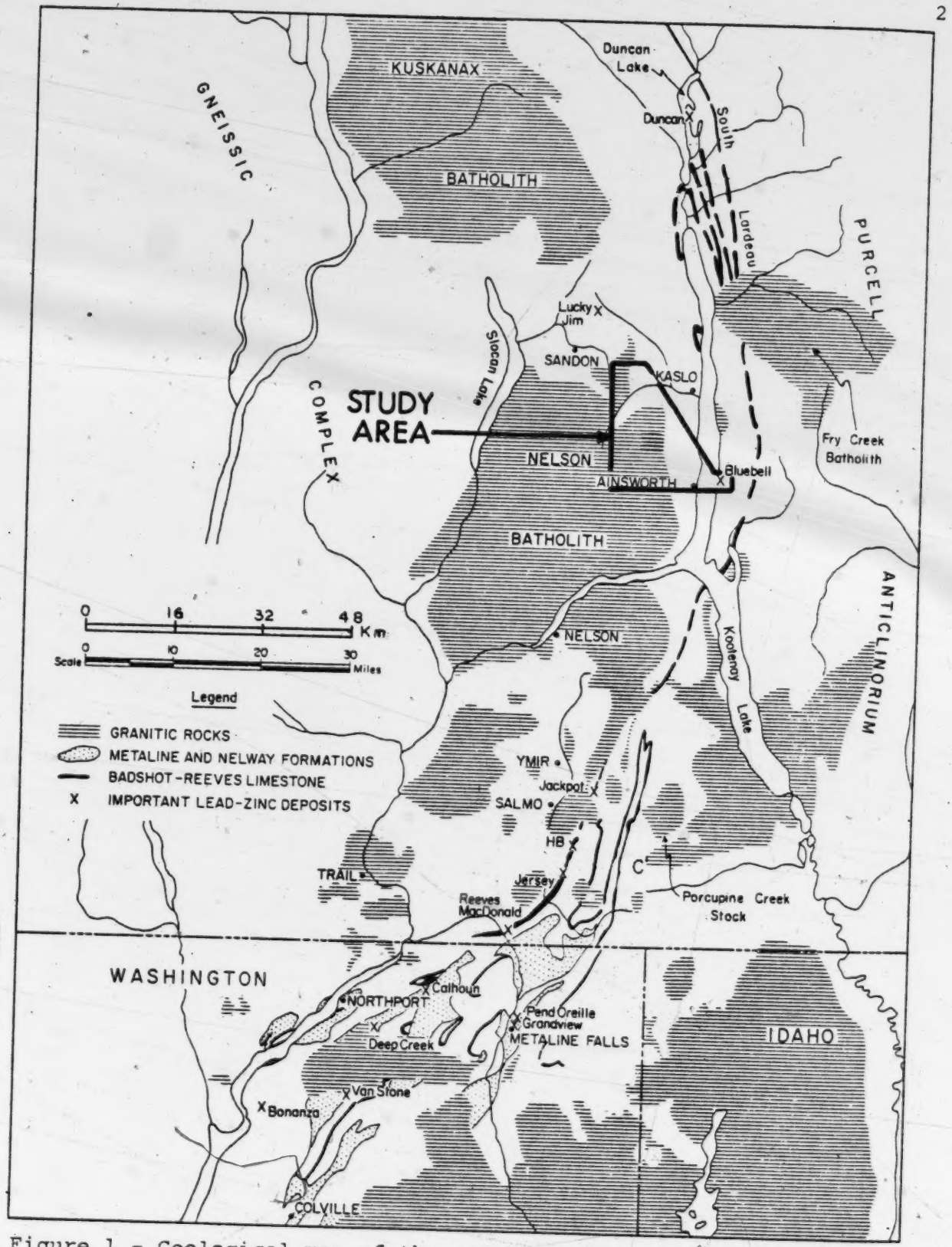


Figure 1 - Geological map of the southern part of the Kootenay Arc showing the location of the field area(adapted from Fyles (1970).

period of time the precise age of which is unknown. Ross (1970) has suggested that the Phase 1 deformation is probably of Permo-Triassic age and associated with lower greenschist facies metamorphism, whereas Phases 2 and 3 are associated with higher metamorphic grades and probably terminated in the mid Jurassic. Superimposed on this earlier folding are local zones of intense deformation adjacent to the batholithic intrusions. The structure of the Kootenay Arc has been documented by Fyles (1967), Crosby (1968), Ross and Kellerhals (1968), Ross (1970) and Hoy (1977).

Coinciding with the structural deformation the Kootenay Arc has been regionally metamorphosed forming a linear metamorphic belt centred along the Purcell Trench and Kootenay Lake. This metamorphism, again prolonged, parallels the structural trends but is locally cross-cutting. Fyles (1967) and Crosby (1968) have indicated that kyanite and sillimanite grades are attained in the vicinity of Kootenay Lake with the grades decreasing rapidly to the west and east. Winzer (1973), however, after a detailed study in the central Kootenay Lake area failed to find cordierite, sillimanite and kyanite on a regional scale. He showed that high grade index minerals are only present in the aureoles of intrusive bodies. The

problems involved in interpreting the metamorphism within the area are numerous including polymetamorphism and retrogression, variation in the lithologies present and non-attainment of equilibrium assemblages.

The Kootenay Arc is part of the southern extension of the Omineca Crystalline Belt in southeast British Columbia. Cox (1979) has postulated an eastward dipping Benioff zone in late Mesozoic times with the Kootenay Arc being marginal between an oceanic environment to the west and a continental crustal environment to the east. Griffiths (1977) has provided a plate tectonic synthesis for the whole of southern British Columbia related to volcanism, plutonism and mineralisation.

CHAPTER II

GEOLOGY OF THE STUDY AREA

a) The Nelson Batholith

The study area (see Figure 2 and Plate I) includes part of the northern and northeastern margin of the Nelson Batholith, a large composite pluton with an areal extent of about 2,000 square km. Specimens from Keen Creek, Long Creek, Flint, Scranton and Woodbury Creek have been studied to give a better understanding of the batholithic rocks in the area (see Plate II). The batholithic host rocks at the Scranton Mine will be described separately in Chapter V.

The main (i.e. non-marginal) phase of the batholith varies between a leuco-quartz monzonite and a granite both of which are distinctly porphyritic. Phenocrysts, up to 5cm in length, but usually about 2cm, are distinctive of the study area and usually consist of microcline microperthites (more rarely sodic orthoclase). These phenocrysts are usually distinctly poikilitic and also strongly zoned. In hand specimen the phenocrysts are white to flesh-coloured but underground, at Scranton, they appear to be more orange in colour. Oligoclase phenocrysts are also present in lesser amounts and typically show carlsbad- and albite-

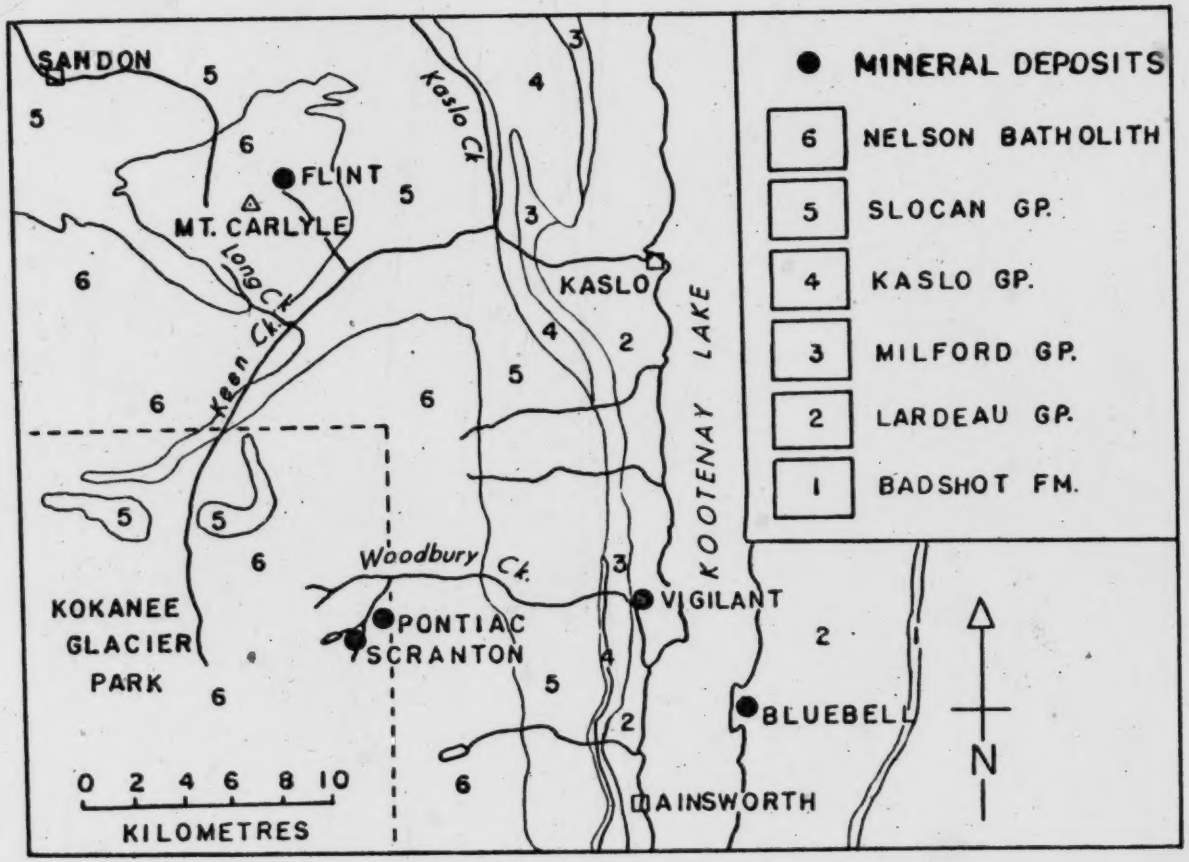


Figure 2- Geological map of the study area

Plate 1

1) Looking southeast upstream along Keen Creek, towards the Kokanee Glacier, taken from the track to Flint. The steep sided slopes are heavily wooded with coniferous stands and thick undergrowth. The geological outcrop here is limited to occasional streams and track sections. Above the tree line steeply cliffed peaks (up to 2000m elevation)provide more continuous outcrop.

2) The discordant contact between the Slocan sediments and the Nelson Batholith exposed on the track to Flint. Compared to the main phase of the batholith the marginal granodiorite here shows a distinct decrease in the size, but not the number, of phenocrysts. The hornfelsed argillites show small scale isoclinal folding that can be related to the regional structural pattern.

PLATE I

Plate II

The Nelson Batholith

1) Photograph of Specimen 35.2, a distincly porphyritic hornblende-quartz-diorite, from upper Woodbury Creek.

2) Photograph of a hand specimen from the 5700 Level, Scranton Mine, showing a distinct contact between a hornblendebiotite rich zone from a felsic zone. Large (up to 5cm) orange coloured microcline (in thin section show a distinct microperthite texture) phenocrysts grow across this boundary suggesting a late stage origin for the K-feldspar phenocrysts.

Photographs of petrographic thin sections showing typical textures

3) Myrmekitic intergrowths, microcline cross hatch twinning (Specimen 19.4 from Long Creek, 2.5x,crossed-nichols.)

4) Extensive Sericitic alteration of plagioclase feldspar (oligoclase) cross cut by late stage carbonate veining (Specimen 13.6 from the track to Flint, 2.5x, crossed nichols).

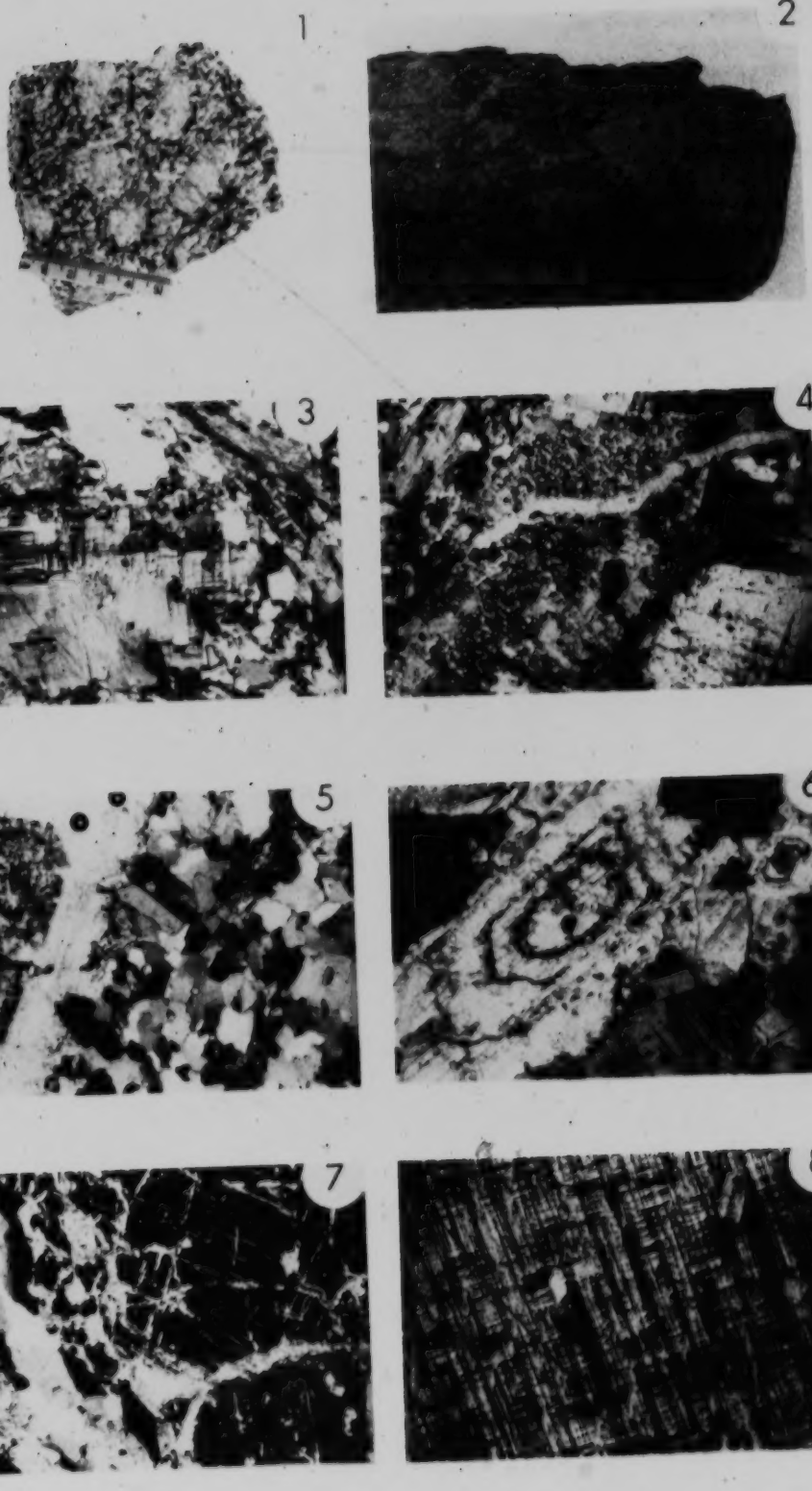
5) This specimen from Long Creek (19.7) is very atypical of the Nelson plutonic rocks in the study area. Petrographic investigation shows it to be an arenaceous metasediment that has been impregnated by the granitic melt. The rock contains abundant recrystallised quartz (approximately 40%) with typically very strongly zoned oligoclase phenocrysts. (2.5x, crossed nichols).

6) A strongly corroded, zoned oligoclase phenocryst from an altered porphyritic quartz monzonite from Flint. The euhedral plagioclase crystals are apparently early in the depositional sequence unlike the K-feldspar (Specimen FI, 2.5x, crossed nichols).

7) Carbonate alteration preferentially along cracks and cleavage planes within microcline microperthite phenocrysts. The specimen (from 5700 level, Scranton Mine) is a porphyritic granodiorite that has been strongly sheared and shows recrystallisation textures (Specimen 5700L, 2.5x, crossed nichols).

8) Distinctly poikilitic texture within a large (5cm) microcline phenocryst from upper Keen Creek (Specimen 17.1, 2.5x, crossed nichols).

PLATE II



twinning. The groundmass shows a hypidiomorphic granular texture composed mainly of quartz, K-feldspar and minor plagioclase (oligoclase). Myrmekitic textures are extremely common where microcline borders plagioclase, producing vermicular quartz crystals. In the main phase rocks the mafic content is usually low (< 6%) with distinctly dark brown to almost colourless pleochroic biotite much more common than hornblende. A large range of accessories are present in all phases of the batholith consisting usually of zircon, epidote and sphene and less commonly apatite, magnetite and rutile.

In many instances the quartz crystals appear distinctly granular showing undulose extinction indicative of shearing and recrystallisation. In more extensively sheared zones (eg. at Flint and Scranton) small granulated quartz crystals have been recrystallised into quartz eyes. Most of the plutonic rocks investigated within the study area are from the marginal phase of the batholith. These are typically very varied, belonging to the hornblende-biotite-diorite granodiorite suite of rocks. Towards the margins as the quartz content decreases there is a corresponding increase in hornblende, biotite and plagioclase (usually slightly more calcic). Using trend surface analyses Sinclair and Libby (1967)

have shown a systemmatic relationship between plagioclase and K-feldspar towards the margin of the Carlyle stock. Quartz diorites and granodiorites are the most common border rocks with often up to 25% mafics, consisting of strongly pleochroic hornblende and biotite. The border rocks are again distinctly porphyritic with zoned oligoclase phenocrysts up to about 1.5cm in length. Microcline commonly occurs but is subordinate to the plagioclase. Poikilitic and myrmekitic textures are both less common in the border rocks compared to the main phase of the batholith. Minor microdiorite inclusions are found within the border phase with relative small (< 1cm) oligoclase phenocrysts in a medium grained hornblende rich matrix. Occasional small (approximately 1cm) quartz phenocrysts, showing distinctly corroded margins, occur in most border rocks.

In upper Woodbury Creek, 2.9km east of the Scranton Mine, a very distinct marginal phase is observed. This has previously been termed a "hornblendite" (Fyles 1967) occurring as "lenticular dyke like masses" within a hornblende microdiorite and quartz diorite. This hornblendite contains a few large phenocrysts of a clinopyroxene, slightly altered to amphibole, which usually only occurs in minor amounts or is totally absent in the border rocks. The augite (?)

phenocrysts show a very pronounced corrosion rim adjacent to the plagioclase rich groundmass. Biotite forms about 10% of the rock and is partly altered to chlorite.

In Long Creek most of the exposed rocks are typical of the marginal phase. However, about 2km from Keen Creek a specimen, that appeared in the field to be a white weathering granodiorite, was found to be an arenaceous metasediment that has been soaked in a granitic melt. The rock consists of strongly corroded and zoned oligoclase phenocrysts in a recrystallised quartz rich groundmass. Elsewhere, Sinclair and Libby (1967) have indicated the presence of a thin sedimentary septum separating the Mt. Carlyle stock from the main phase of the Nelson Batholith to the south. No evidence was found for a septum "several hundred feet wide." The rocks in Long Creek are, however, extensively altered and recrystallised indicating that a shear zone separates the stock to the north from the main plutonic body.

Alteration, often extensive, of the coarse grained plutonic rocks is very common. Preferential alteration of the central (more sodic rich) portion of zoned phenocrysts to sericite typically occurs. Titaniferous ilmenite has altered to give abundant

sphene and chlorite has occasionally been produced at the expense of hornblende and biotite. In places (eg. Flint) late stage carbonate (calcite and siderite) forms small cross cutting veinlets and also extensive pervasion of the rocks. The carbonate alteration is distinctly later than that of the sericite.

The rocks in the study area are generally massive and unfoliated. In places, however, expecially near the margin, slight orientation of the mafics is visible which is attributed to flowage rather than deformation since these rocks are not more crushed or strained than elsewhere.

Origin of the Nelson Batholith

Gabrielse and Reesor (1964) have documented values for the age of various phases of the composite Nelson Batholith. The Mt.Carlyle stock gives a biotite K/Ar date of 171Ma for the marginal phase. The autochthonous core of the batholith has been dated by Baadsgaard et al (1961) at 95Ma (mid Cretaceous) with an updating episode related to the proximity of the Coryell Tertiary syenite intrusives at 50 - 60Ma. More recent work by Nguyen et al (1968), using biotite and hornblende K/Ar dates have confirmed a minimum age of 158$\pm$ 7Ma for plutonic rocks at the northern margin of the batholith. These ages indicate that the marginal phases of the batholith are the oldest

being emplaced at approximately the end of the Middle Jurassic.

The crystallisation history of the plutonic rocks is obviously very complex with small stocks and plutons coalescing to form the large composite batholith. This complexity is further suggested by the nature of the distinctly zoned phenocrysts. Crosby (1968) suggests that late-stage solutions from the still liquid interior might be responsible for the addition of quartz and microcline to the recrystallised plutonic rocks, especially in the border zone. At Flint the texture in a porphyritic granite suggests a multi stage history with euhedral oligoclase phenocrysts apparently earlier and microcline micro-perthite phenocrysts later than the bulk of the groundmass. The K-feldspars were apparently formed, therefore, after emplacement of the batholith, probably by later metasomatic deep seated alkaline solutions.

Seven plutonic rock samples were analysed by X.R.F. in order that a further understanding of their geochemistry and origin could be interpreted. Appendix 1 documents the values and provides a key for the specimens analysed. Variation diagrams have been plotted using part of this XRF data with Figure 3 showing an AFM plot; Figure 4 an MgO/SiO_2

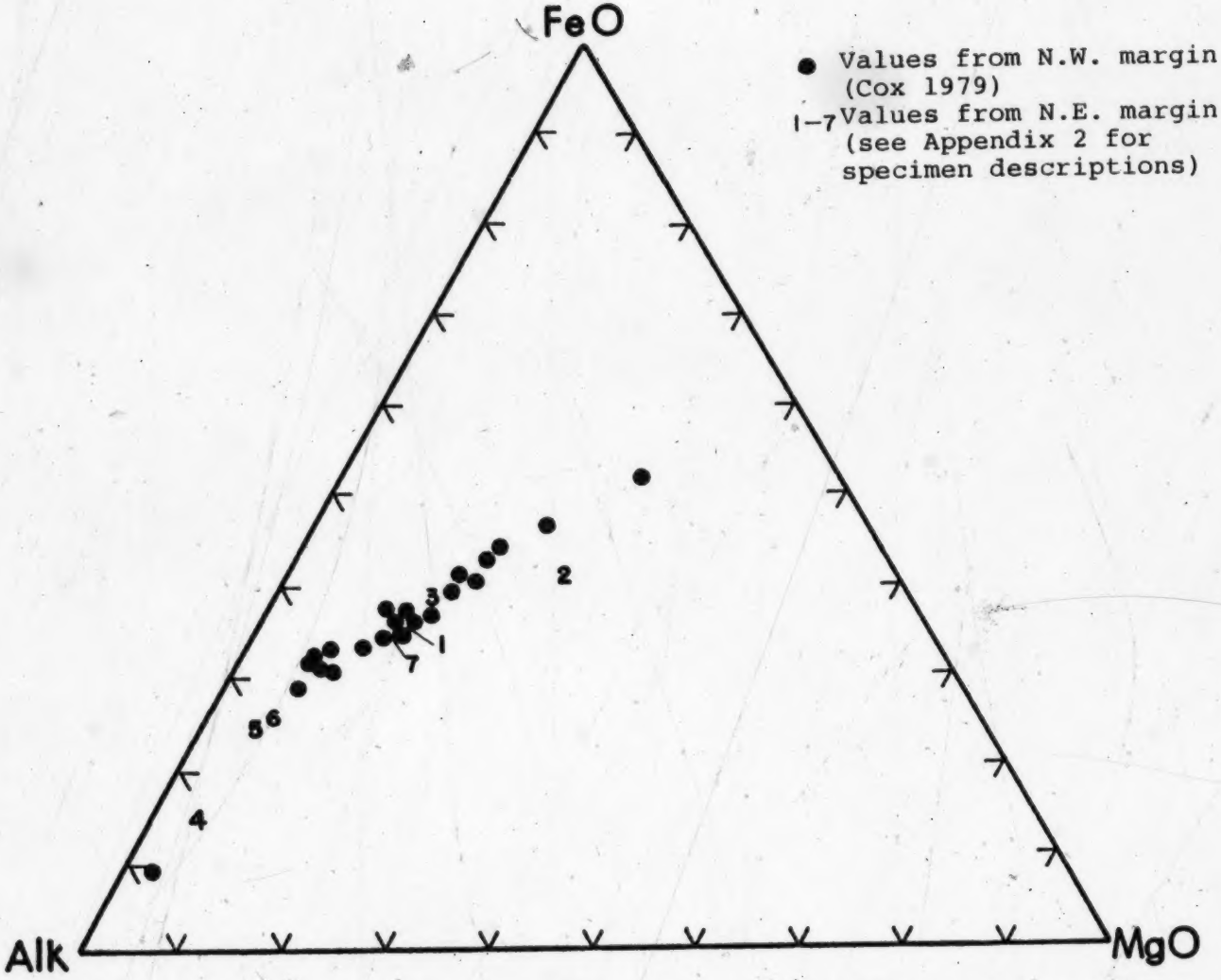


Figure 3 - AFM plot of the plutonic rocks from the northern margin of the Nelson Batholith.

Figure 4 - MgO/SiO_2 Larsen diagram of plutonic rocks from the northern margin of the Nelson Batholith.

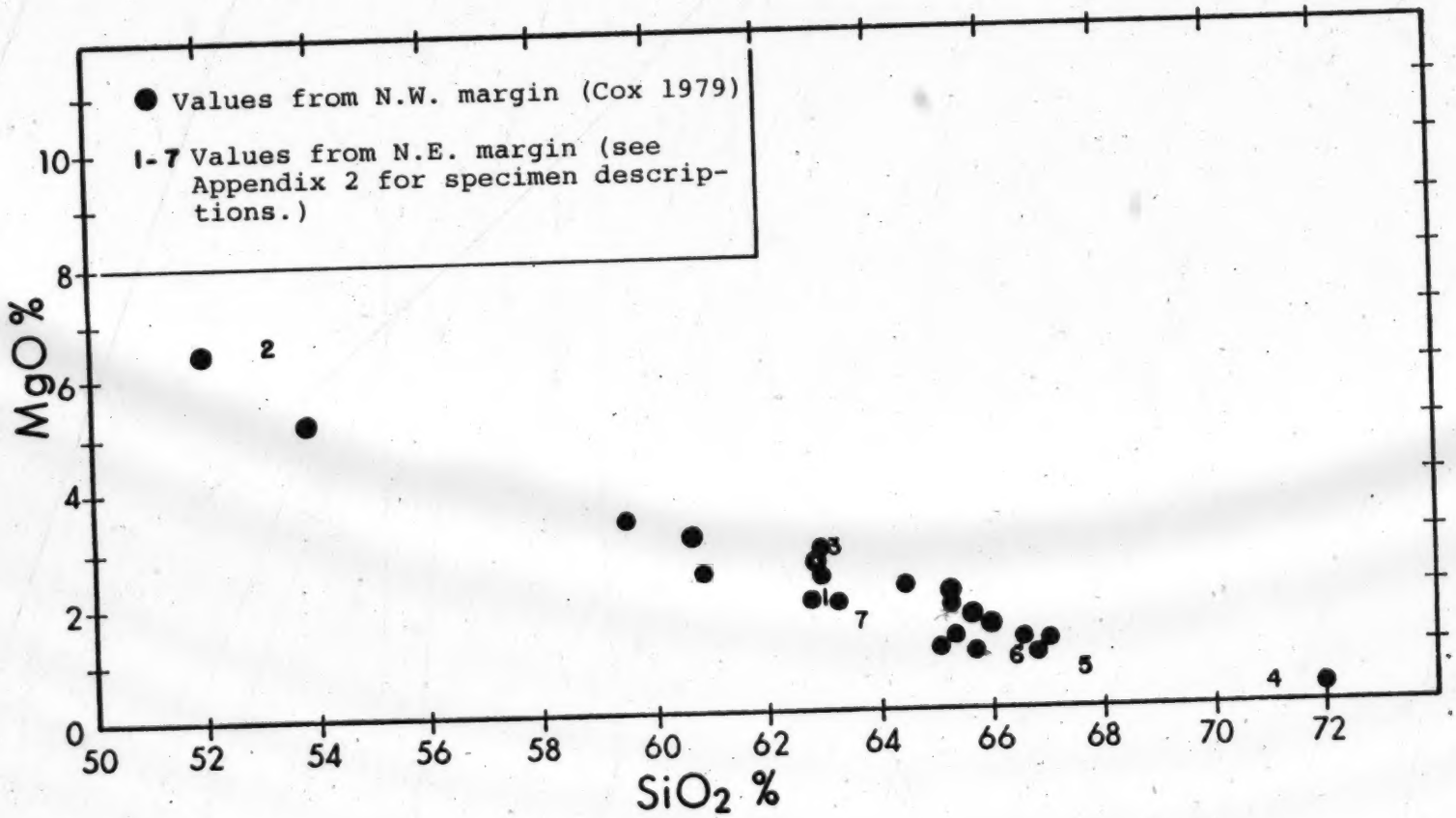
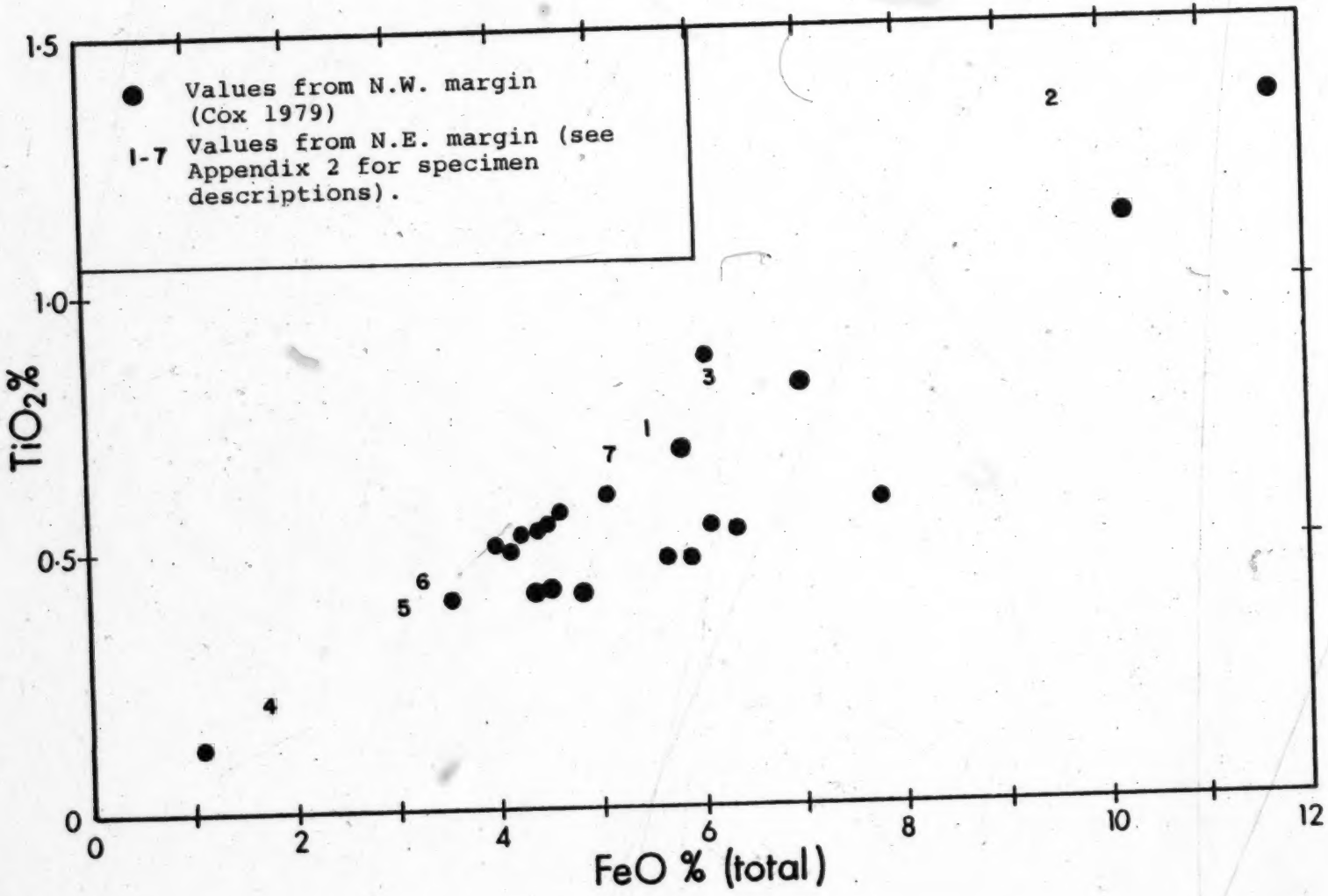


Figure 5 - TiO_2/FeO plot of plutonic rocks from the northern margin of the Nelson Batholith.



Larsen diagram and Figure 5 a TiO_2/FeO plot. Values provided by Cox (1979), from the northern margin of the Nelson Batholith, were also plotted so that a direct comparison can be made. It can be seen that all the values plotted, for both the northern and northeastern margins, follow distinct calc-alkaline trends, with apparently a similar crystallisation history. The presence of the Rossland Volcanics, also distinctly calc-alkaline in character (Beddoe-Stevens 1977), further to the south, has been discussed by Cox (1979) who has interpreted an eastward dipping Benioff zone in late Mesozoic times. The Nelson Batholith occurs to the east of Moore's (1959) "quartz diorite line" where the Moho is deeper as a result of thickened continental crust. This therefore implies an increase in the overall silica and potassium contents of the rocks due to the crustal influence during fractional crystallisation of the calc-alkaline magmas. Whole rock $^{87}Sr/^{86}Sr$ values of approximately 0.707 indicate an upper mantle or base of crust origin for the magmas forming the Nelson Batholith which have become only slightly enriched in ^{87}Sr during their upward migration (Fairbairn et al 1964). The Nelson Batholith appears to be distinctly"I type"in character

(White and Chappell 1977) resulting from partial melting from an igneous source, again probably the upper mantle. This is partly shown by the presence of hornblende, SiO_2 contents between 55% and 75% (suggesting crystal fractionation has occurred) and an orderly Larsen diagram plot (eg. MgO/SiO_2). The presence of cognate inclusions (ie. solidified equivalents of cogenetic magmas from earlier in the descent) of basic character is also indicative of I type granites.

The Nelson Batholith appears to belong to Buddington's (1959) "mesozonal" classification, having a complex emplacement history, being in part discordant and in part concordant. The core of the batholith is believed to be autochthonous (Little 1960) being formed in situ by granitisation. The margins, however, show distinctly mechanical intrusion involving mobilisation and injection, with vertical upward flow of magma. Gabrielse and Reesor (1974) have documented that the Nelson Batholith is a "northern plunging wedged shaped mass with a bulbous upper zone and a narrow gneissic keel". The presence of wedged shaped apophyses at the margins of the batholith, poorly developed internal vertical foliation and lineation and the discordant nature all suggest that the batholith was intruded by dyking and stoping.

b) The Sedimentary Succession

Within the study area the rocks range from Cambrian to Lower Jurassic, being part of the Central Kootenay Arc succession described by Crosby (1968). Fyles (1967) has stated that there is no stratigraphy, in the strictest sense, along the west side of Kootenay Lake "merely a succession of lenticular units that are the deformed remnants of what once was a more or less continuous stratigraphic succession." However, a generalised stratigraphy can be interpreted which including the succession to the east of Kootenay Lake, provides a full cross section across the Kootenay Arc in this area.

i) Badshot Formation

This formation, which was used by Fyles (1962) to define the eastern margin of the Kootenay Arc, is locally called the Bluebell Limestone at Riondel. It is a white to creamy coloured dolomitic marble or coarse-grained calc-silicate rock of variable thickness (15 to 100m averaging about 20m). Quartz is commonly present with a number of other accessory calc-silicate minerals (eg. diopside, tremolite, phlogopite). Archaeocyathids indicate a Lower Cambrian age.

ii) Lardeau Group

The Lardeau Group, directly overlying the Badshot Formation (Hoy 1977), flanks both the east

and west sides of Kootenay Lake in the study area and apparently thickens northwards. The Lardeau Group is subdivided on a strictly lithologic basis into three units (Crosby 1968). Because of numerous sedimentary facies changes, low to high grade regional metamorphism and structural complexities in the area only local lithologic correlations are possible. The lowermost unit can be subdivided into the Early Bird Formation and the Princess Formation (Fyles 1967) consisting mainly of micaceous schists, micaceous quartzites and marbles. The middle and upper units, almost entirely confined to the eastern side of Kootenay Lake, consists mainly of massive and laminated calc-silicate gneisses and feldspar-quartz-biotite-muscovite gneisses with alternating micaceous and quartzofeldspathic layers.

iii) Milford Group

This group, originally classified by Schofield (1920) as the Ainsworth and Josephine Formations, consists of several thousand metres of black argillites and slates with limestone and chert interbeds in the type section. In the study area, however, the Milford Group is very much thinner (less than 70m) and the overlying Kaslo Group rests directly on the basal part of the Ainsworth Formation. A basal conglomerate occurs locally overlain to the north of Ainsworth by a dark grey and white, laminated or mottled fine-grained limestone. Amphibolites and

biotite schists are interpreted as sill-like intrusives of the overlying Kaslo Group. The base of the Ainsworth Formation is thought to be Mississippian in age with the top of the Milford Group (not represented in the study area) probably of Triassic age (Crosby 1968).

iv) Kaslo Group

To the north of the Kaslo River the Kaslo Group is about 1000m thick but rapidly thins towards the south. It overlies the Milford Group unconformably and is overlain disconformably to the west by the Slocan Group. The Kaslo Group is distinctly volcanic in origin being predominantly volcanic breccias and tuffs at the base of the sequence, with andesitic and dacitic flows and sill like bodies in the upper part. Any primary structures, however, have been almost completely obliterated by metamorphic recrystallisation and intense deformation. As a result the sequence is now represented by monotonous fine-grained laminated amphibolites and chlorite-biotite schists with some greenstones and pyroclastics. The Kaslo Group is probably Triassic in age (Little 1960).

v) Slocan Group

Because of structural complexities, the various estimates for the thickness of the group range from

approximately 2000m (Cairnes 1934) to 10,000m (Irwin 1951). Repetition of beds produced by large scale recumbent folding indicates that the lower estimate is probably more correct. The "Slocan Fold" (Hedley 1945) is a large recumbent syncline facing southwest with horizontal axial plane and cleavages. The Slocan Group consists mainly of argillaceous rocks but fine-grained quartzites and limestones are abundant particularly in the middle of the sequence (eg. upper Keen Creek). Cox (1979) has interpreted the succession, which shows graded bedding and soft sediment deformation, as a distal turbidite sequence. He has shown that the lower part of the Slocan Group has been derived predominantly from a continental crustal environment (probably in the east) but the upper part is more basic in character. This has been correlated with the formation, uplift and erosion of the Rossland Volcanic Group which according to McConnell and Brock (1904) overlies the Slocan Group to the northwest of Slocan Lake. The lower part of the Slocan Group is interpreted as being Triassic with the top Lower Jurassic in age.

Most of the Slocan Group is in the biotite grade of regional metamorphism, but a zone of deformation and thermal metamorphism related to the Nelson

Batholith extends as much as 800m east and 1500m north from the exposed contact. This thermal effect is superimposed on the regional deformation and metamorphism. In Keen Creek the sediments are strongly foliated and contain biotite, epidote, quartz with minor sphene and cordierite. More usually in the highly carbonaceous black fissile argillaceous sediments andalusite megacrysts occur which may be rimmed by sillimanite (Fyles 1967). Also within the contact aureole thin tremolite-and diopside-bearing marbles occur.

CHAPTER III

LEAD ZINC DEPOSITS OF THE KOOTENAY ARC

The area of southeast British Columbia and adjacent parts of northern Washinton, Montana and Idaho is recognised as the most important PbZn metallogenic province of the Cordilleran region (Fyles 1967). This metallogenic province contains the Sullivan orebody at Kimberley, the deposits of the Shuswap Metamorphic Complex and East Kootenays, the Coeur d'Alene district of Idaho and also PbZn deposits within the Kootenay Arc itself.

Within the Kootenay Arc there is a large variety of PbZn deposits, which can roughly be described as Mississippi Valley Type deposits. Figure 1(modified from Fyles 1970) shows the main deposits, including the Slocan and Ainsworth deposits, Bluebell, Salmo and the Metaline Falls district in northeast Washington. Fyles (1966) has subdivided the PbZn deposits into two main categories i.e. those concordant or those transgressive in type. The Salmo, Metaline and Shuswap deposits all belong to the concordant type whereas the Slocan, Ainsworth and Bluebell deposits are classified as transgressive.

a) Concordant Deposits

The concordant deposits, usually the largest and most productive, have simple mineral assemblages of pyrite, sphalerite and galena with or without pyrrhotite. The sulphides usually occur as disseminations in a calcareous host rock with a gangue of dolomite, calcite, quartz and rare barite. The mineralisation appears to be locally controlled by the structure (for example in the Salmo type deposits by tight folding and in the Metaline district by faulting and brecciation.) Most of these deposits have been affected by the regional metamorphism (Muraro 1966) which makes them distinctly older than the transgressive type.

b) Transgressive Deposits

These deposits have essentially been localised by shear zones, fracture zones and faults which cut across a variety of rock types. These deposits can be subdivided into replacement deposits in which fissure filling is relatively subordinate (eg. Bluebell) and into vein deposits in which replacement is subordinate (eg. Slocan). Their mineralogy is more complex than those of the concordant types but generally is still relatively simple. Galena, sphalerite, pyrite and pyrrhotite are still the most common minerals but chalcopyrite, argentiferous tetrahedrite and other silver-bearing minerals also occur. The age of this type of

mineralisation is generally post-regional metamorphism with the source of the solutions apparently more localised than that of the concordant deposits.

i) The Bluebell Mine

The Bluebell mine lies on the east shore of Kootenay Lake almost directly east, across the lake from the Ainsworth camp. Before its closure, in 1971, it was one of the major PbZn ore deposits of British Columbia. The stratigraphy on the east side of Kootenay Lake has been summarised by Ohmoto and Rye (1970). The beds dip westerly between 15° and 55° with the sequence being overturned (Fyles 1967). Hoy (1977) has shown the structure to be a recumbent anticline which he called the "Riondel Nappe". The Bluebell deposit is located within the high-grade regionally metamorphosed sector of the central Kootenay Arc.

An outline of the geology of the Bluebell Mine is provided by Ransom (1977). The orebodies are essentially massive replacements of the Bluebell Limestone, locally between 30 and 50 metres thick, which is correlated with the Lower Cambrian (Lardeau) Badshot Limestone. The hanging wall rocks consist of mica schists and micaceous quartzites with calcareous lenses (eg. the Upper Limestone). The footwall rocks are calcareous quartz mica schists and hornblende schists.

The ore deposits consist of three main ore zones, spaced at about 500 metre intervals along the strike of the Bluebell Limestone. The mineralisation appears to be fracture controlled with sulphide replacement extending laterally from steep cross fractures. Irvine (1957) has described the ore zones in detail and has shown that the most favourable zone of replacement is in the dense, closely-banded limestone horizons of the upper Bluebell Limestone. Ohmoto and Rye (1970) have postulated a paragenetic history of mineral deposition at Bluebell:

Period I - Formation of knebelite, (a high temperature FeMn olivine.)

Period II - Deposition of the main massive sulphide-quartz-carbonate ores. The generally coarse-grained sulphides consist mainly of pyrrhotite, sphalerite and galena with minor but ubiquitous pyrite, chalcopyrite and arsenopyrite. Temperatures exceeded 450°C.

Period III - Development of free standing crystals of quartz, calcite and sulphides in vugs. Temperatures indicated by fluid inclusions are 450 - 320°C.

The mineralisation cross-cuts a series of lamprophyre dykes which have recently been indicated to be Cenozoic in age (Fyles 1967) from field relationships. As yet, however, the age of the lamprophyre dykes has not been asertained and as a result the age of the

mineralisation is still unknown. (P. Ransom, Cominco Geologist personal communication).

ii) Ainsworth

The initial work on the mineralisation in the Ainsworth district was undertaken by Schofield (1920) and Rice (1944) and has most recently been documented by Fyles (1967).

Fissure veining and replacement are both quite common, depending upon the wall rock type. Fissure veining occurs predominantly in mica and hornblende schists and quartzites with replacement in more calcareous beds. The mineralisation is generally confined to the Milford and Lardeau groups, within the biotite-garnet grade of regional metamorphism. The individual orebodies are small, irregular in shape, but generally high grade in nature. The width of the deposits varies from a few centimetres to a metre or so and rarely exceeds 100 metres down dip or along strike.

The major sulphide minerals are pyrite, galena, sphalerite and pyrrhotite, while chalcopyrite and arsenopyrite are minor constituents. Pyrrhotite is usually rimmed or replaced by pyrite along fractures. The pyrrhotite suggests a higher temperature origin of the ore but is generally separated in time from the

other sulphides. The gangue is usually quartz, calcite, siderite and occasionally minor fluorite. The wall rock alteration contains chlorite, sericite and carbonate and extends for less than one metre from the fissures. Knebelite, associated with pyrrhotite, occurs rarely in the northern part of the Ainsworth camp. The mineralisation appears to be controlled by a simple set of fractures which have undergone repeated movements. These fractures belong to the same set that controls the Bluebell mineralisation. The variation in the degree of mineralisation is perhaps controlled by the fracture intensity and spacing and as such Ainsworth, having only minor fracturing, is less mineralised than the Bluebell deposits. The presence of knebelite and pyrrhotite in the northern part of the Ainsworth camp, however, shows that a distinct parallel can be drawn with Bluebell. As at the Bluebell Mine, the Ainsworth mineralisation has tentatively been dated as Cenozoic in age.

iii) Slocan

Lindgren (1919) classified the Slocan deposits as "typically mesothermal in character." Since that time much work has been done on the mineralisation with two important papers by Cairnes (1934)

and Hedley (1952).

The host rocks for the mineralisation are the Triassic to Lower Jurassic Slocan sediments (eg. the Silmonac Mine at Sandon) and to a minor extent the Nelson Batholith (eg. Scranton). The mineralisation is usually distinctly transgressive, occurring in shear veins within the slaty fissile argillites, and with the hydrothermal solutions apparently having access to all cross cutting fractures. The deposits vary widely in size, outline and continuity with very little replacement, despite the presence of some limestone in the sequence.

A large number of minerals have been recognised in the Slocan deposits with sphalerite and galena dominant and pyrite, chalcopyrite, pyrrhotite and arsenopyrite also present. Silver is found in the form of freibergite, argentite, pyrargyrite and native silver. The gangue is usually carbonate-rich with calcite, siderite and minor quartz. A common feature of the Slocan ores is the presence of gneissic galena (Uglow 1917).

CHAPTER IV

HISTORY OF EXPLORATION AND DEVELOPMENT AT SCRANTON MINE

The Scranton property comprises 11 claims situated near the head of Pontiac Creek, a northerly flowing tributary of Woodbury Creek, 22 miles northeast of Nelson. The claims lie within the Kokanee Glacier Provincial Park, about 1km from the eastern boundary, between an elevation of 1,900m and 2,300m (see Figure 6)

A series of northeast-striking quartz veins in the porphyritic granite are exposed at intervals for a distance of about 2,500m. Most of the early exploration work on this vein system was done at its extremities, with an initial production in 1898 from the Pontiac (northeast) and Tecumsie (southwest) claims. Cairnes (1934) states that the first recorded production was 20 tons of ore averaging 50oz. silver per ton and 30% lead. During the period 1900-1904, underground development and exploration was done over a length of 120m and a depth of 60m in 3 adits. Ore shipments were made intermittently until 1905 but, except for minor development, the property remained idle until 1930. Cairnes (1934) reports that the workings were "obscured" in the summer of 1927.

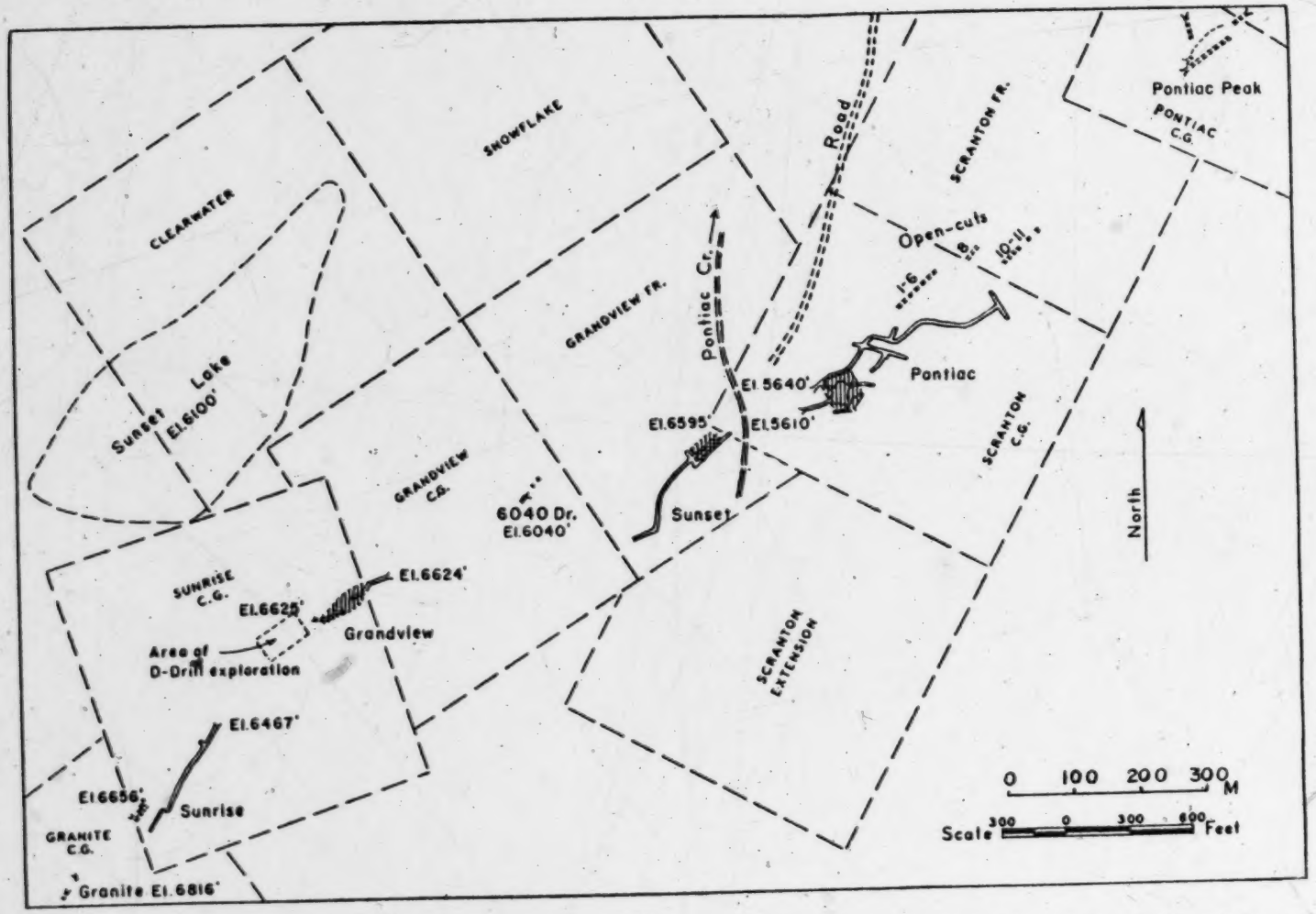


Figure 6 - Property and workings, Scranton Mine (adapted from British Columbia Department of Mines Annual Report 1954 p. A 135).

In 1930 the Pontiac, Scranton and Sunrise claim groups were acquired by the Scranton Consolidated Mining Company Ltd. Extensive development work was carried out by them at the lower elevations in the midsection of the property, especially on the Scranton (Lower Pontiac) workings on the east side of Pontiac Creek. In 1939-1940 the main workings on the Scranton claim were driven. A private road, linking the mine camp to the Nelson-Kaslo Highway, 8 miles south of Kaslo, was started in 1945 and completed in 1947.

Little (1960) states that the total production of ore from the property from 1898 to 1954 inclusive was 6,881 tons containing 1,426,150 lbs. lead; 1,108,852 lbs. zinc; 73,227 oz. silver; 1,359 oz. gold and 5,539 lbs. cadmium. This represents a calculated average grade of ore: gold 0.22 oz. per ton; silver 10 oz. per ton; lead 11.9% and zinc 10.6%. This production was probably derived from both the Scranton and the Sunset drift adits (on the west bank of Pontiac Creek on the Grandview claim.)

The name of the company was changed to Scranton Mines Ltd. in 1951 and in 1953 the new company drilled 11 holes totalling 600m to determine the vein extension on the Sunrise claim. By the end of 1952 the workings

comprised about 300m of drifts and crosscuts and a 25m winze.

Little further activity took place until 1960-61 when an additional 30m of drifting, crosscutting and underground drilling was carried out in the Scranton workings. Blue Star Mines Ltd. optioned the property in 1964 and carried out stripping and diamond drilling on the Scranton claim.

No further activity was reported until 1967 when exploration was begun on a new adit at the 5,900 foot (1,798m) elevation on the southeast edge of the Grandview claim ("West Sunset" workings). The adit was driven as a crosscut for 30m, intersecting a thick quartz vein which was drifted along.

Silver Star Mines Ltd. acquired the property in 1967 by a stock transaction and they began a new adit at the 5,700 foot (1,738m) level to explore the downward extension of mineralisation in the 5,900 foot (1,798m)elevation level; drifting, raising and stoping were continued at both levels. Stoping of ore proved uneconomical and the operation was shut down in 1970. Adit, trench and drill indicated reserves on the 5,700 and 5,900 levels (West Sunset workings) were estimated

in November, 1969 by W.M. Sharp at 47,245 tons of ore averaging 0.235 oz. gold and 7.0 oz. silver per ton; 9.1% lead; 7.6% zinc, plus minor cadmium values (Northern Miner: March 12, 1970).

David Minerals Ltd. acquired the property from Silver Star Mines, and the Ainsworth Mill from Blue Star Mines, in 1977. They optioned the property to the present mine operator, HEM Mines Ltd., on a short-term mining lease. HEM Mines have been working the property since July, 1977. The Sunrise Basin - Grandview area has been given their immediate interest and as a result they have undertaken a line drive, which was started in mid March, 1978, entering the 5,700 foot (1,738m) level extending southwards for about 300m from the portal. "The line drive was planned to explore the ore-making potential of the downward extension of mineralisation located over an 800 foot (244m) length on surface and in a series of 5 surface diamond drill holes in what is called the Sunrise Basin and Grandview areas of the mine." (David Minerals Ltd. Letter to the Shareholders, June 28, 1978). The drift encountered mineralisation between 490 feet (149m) and 570 feet (174m) from the start of the drive. The best grab samples from this area returned 38 oz. silver per ton plus approximately 0.1 oz. gold per ton. The operators intend to continue

drifting along the mineralised structure and drive crosscuts every 30m for drill stations. At each station fans will be drilled off to check the mineralisation above and below the level. The drilling will be the next step in establishing a series of ore shoots in the ore zone. As a result of this continued exploration, an intersection of about 70cm of massive sulphides, in the drift at approximately 230m from the start of the new drift, has been found. This section assayed 0.22 oz. gold per ton; 3.57 oz. silver per ton; 3.9% lead and 1.3% zinc. Additional exploration work was undertaken later in the 1978 season on two other zones, the west Sunrise and the southwest Sunrise, with again the immediate objective that of detailing of ore shoots within the shear system. British Columbia Department of Mines Annual Reports show that the property has,up to 1977, an estimated total production of 17,600 tons grading: 0.16 oz. gold per ton; 6.2 oz. silver per ton; 6.8% lead; 5.5% zinc and 0.08% cadmium.

The present operators are trucking the ore to their Ainsworth mill where it is crushed to 100 mesh then, using a flotation method, the galena and sphalerite concentrates are separated from the pyrite. 90% of the

gold is recovered, with only 10% remaining in the pyrite tailings. The concentrated values (see Table 1) are approximately 10 times higher than those of the mined ore.

Table 1 Sample Assay Returns

November 2nd, 1977 Zinc Concentrate

oz/ton Au	oz/ton Ag	% Pb	% Zn	% S	% SiO_2		
0.405	10.75	2.5	54.7	33.3	1.2		
% Fe	% CaO	% MgO	% Sb	% As	% Cd		
6.7	0.1	0.02	0.03	0.02	1.3		99.87%

March 27th, 1978 Lead Concentrate

oz/ton Au	oz/ton Ag	% Cu	% Pb	% Zn	% S		
2.503	42.45	0.01	71.3	4.5	17.8		
% SiO_2	% Al_2O_3	% Fe	% CaO	% Sb	% As	% Bi	
1.3	0.1	3.9	0.1	0.1	0.02	0.01	99.13%

CHAPTER V

GEOLOGICAL DESCRIPTION OF THE SCRANTON DEPOSIT

Scranton is a lead-silver-gold-zinc-cadmium deposit situated at the northeastern margin of the Nelson Batholith. In the vicinity of the mine the "batholith" phase is a coarse grained porphyritic granodiorite which constitutes the host rock for the mineralisation.

The contact of the batholith with the sediments is located at the extreme northeast portion of the property within the Pontiac workings. The contact cannot be observed on surface due to surficial debris. The mine workings are in disrepair and have been closed since the 1920's. Little (1960) noted that the sediments in contact with the granodiorite are quartzites with a few metasediment xenoliths occuring in the marginal phase. These xenoliths are generally very small although "a huge xenolith of quartzite encased in granite" has been noted in the main Pontiac drift from 100m to 250m from the portal. This could, however, be simply due to an inflection in the batholithic margin.

The mineralisation occurs in a series of veins striking northwesterly and generally dipping steeply to the southwest. The veins occur in a shear zone

10m or more in width, but despite the extent of the shearing there appears to be little or no fault displacement. Diamond drilling on the 1,738m (5700 foot) level 80 metres out from the main shear shows that there are no other parallel fissures. The individual mineralised veins range up to 1m in width and average about 65cm. On surface the vein system has been traced for at least 2,000m but only a few sectors show appreciable mineralisation. Cairnes (1934) in his description of properties, Slocan Mining Camp area, expresses doubt about the continuity of these veins.

The gangue is mainly massive milky quartz but occasionally shows honeycombed and vuggy textures. Although quartz predominates within the veins, carbonate (mainly a white iron poor calcite, and occasionally a little siderite) is also found. In intensely sheared zones brecciated wall rock forms abundant gangue. The shear zone usually carries several quartz veins and more irregular quartz pods.

The veins are locally well mineralised carrying shoots of clear and mixed ore. Galena is widely disseminated and in addition forms streaks and pods of fine-grained and coarse-grained cubic galena. However there is very little or no dissemination of the ore

minerals into the wall rocks. Sphalerite is less noticeable than the galena and is usually fine-grained and pale brown in colour. Fine-grained and cubic pyrite is found in the barren quartz shears as well as associated with the massive sulphides. Very rarely arsenopyrite and chalcopyrite are present. Little (1960) also notes that grey copper, ruby silver, native silver and argentite have been reported from this type of vein, however, none were observed in hand specimen at the Scranton Mine by the writer.

The Pontiac workings are located at the extreme northeast portion of the shear system on the east side of Pontiac Creek. Here there appears to be two main veins, roughly parallel, striking northeast to east but averaging N. 60° E. The dips range from 10° to 70° southwest averaging about 30°. In the granodiorite the veins are up to 60cm wide, containing disseminated sulphides and pockets or small shoots of high-grade sulphides, but pinch out as the contact with the quartzite is approached.

In the Scranton Mine proper (ie. the Sunset workings, which have now been extended further to the southwest) on the west side of Pontiac Creek the vein strikes northeast-southwest and dips at approximately 60° south. The drift follows the persistent fissure system

which is a southwesterly extension of that observed at the Pontiac workings. The initial part of the 1,738m drift follows a persistent quartz-rich shear about 1m wide and mineralised in part with massive cubic galena and minor sphalerite. Large angular granodiorite blocks up to about 80cm in length are caught up in the veins and a thick, up to 10cm,gouge of crushed wall rock forms the contact with the massive porphyritic granodiorite. Between the 1,798m (5,900 foot) and 1,738m (5,700 foot) levels the vein system expands into a rich pod, about 20m long, of massive sulphides, which has been mined out. Past the "ore zone", about 30m from the portal in the 1,738m level (ie. within the Sunrise Basin and Grandview areas of the mine) the vein pinches and splits rapidly into 3 or 4 thin, relatively unmineralised quartz veins with minor galena and pyrite.

Between 1,490 (454m) and 1,570 (479m) feet from the portal in the 1,738m level the new drift encounters more mineralised shearing with high silver and gold values in grab samples. Further into the adit,fracturing becomes more intense and disseminated cubic pyrite is the only sulphide found. Chlorite-and talc-covered slickenslides are present in a 2m wide sheared zone dipping at about 40° south (ie. at a slightly lower

angle than at the portal). Because of the extensive fracturing and alteration of the porphyritic granodiorite here, the mine operators are hoping to encounter more mineralisation as the drift is extended southwards.

On surface the granodiorite phase is usually extensively altered and friable with abundant kaolinisation. Underground, adjacent to the shear zone at the 1,738m level, the country rock is a distinctly porphyritic, coarse-grained granodiorite. Up to 30%, but usually a little less, of the whole rock is made up of K-feldspar phenocrysts. These phenocrysts are generally 3-5cm long microcline microperthites, and are very distinctive of the Nelson granodiorite within the mine area and also on a more regional scale. They often show poikilitic textures in thin section and are occasionally myremekitic where the microcline borders plagioclase. The phenocrysts are usually distinctly pinkish in colour, but further underground they appear more orange, and except for minor sericitic alteration along fractures are relatively fresh and unaltered.

The K-feldspar phenocrysts are found in a hypidiomorphic groundmass consisting mainly of quartz, up to 30% but usually about 15%, and altered plagioclase feldspar. The quartz usually forms augen or clumps of

anhedral, granulated crystals suggesting a history of shearing and recrystallisation. Oligoclase, forming up to 60% of the groundmass, is strongly altered to sericite. The mafic content of the granodiorite can vary greatly,usually about 5% and consists of biotite often altered fully or in part to chlorite. Hornblende is a rare, but noticeable, component of the usual granodiorite underground. A sample taken from the 1,738m level shows a distinct contact between two phases of the granodiorite. A sharp contact separates a rock almost devoid of mafics from a more mafic phase with about 35% biotite and hornblende. An interesting point is that the K-feldspar phenocrysts have formed across the junction, with obvious implications as to their genesis.

A "black mafic pod" is to be found on the main level at the Scranton Mine. This is a quartz micro-diorite with again small feldspar phenocrysts which suggests a strong genetic link with the main granodiorite phase. About 1% of this rock consists of oligoclase phenocrysts up to 0.6cm in length. Hypidiomorphic quartz (28%) and oligoclase (50%) form the bulk of the groundmass with green hornblende (10%) and biotite (10%) the remainder. Accessory sphene and apatite are present in most phases of the granodiorite.

Within the main shear system the granodiorite appears relatively bleached underground and has been mistaken in the past for quartzite. Fine grained carbonate is found within the mineralised shear zone, along cracks and shear planes, and appears to be younger paragenetically than the sericitic alteration. Using the energy dispersive method, E.D.A., on the electron microprobe the carbonate was analysed and found to be $FeMnCO_3$ with minor calcium and magnesium.

Minor euhedral pyrite occurs pervasively, adjacent to the mineralised zones, within the wall rocks.

Plate III shows typical ore from Scranton with Plate IV showing cataclastic and replacement textures commonly present within the ore.

Plate III

1) Photograph of a slab of ore from the Scranton Mine showing a slight banded texture caused probably by several periods of intrusion. The gangue is a massive milky white quartz with very minor carbonate.

2) Sample 5702-5 III showing highly reflective electrum blebs (E) and included sphalerite within pyrite. The concave shape of the sphalerite suggests it has been replaced by the pyrite. (16x, reflected light microscopy, plane polarised light).

3) Sample 5702-5 I showing replacement textures within the ore from Scranton. Euhedral pyrite, distinctly rimming the central sphalerite, appears to be undergoing replacement by the galena. (4x, reflected plane polarised light).

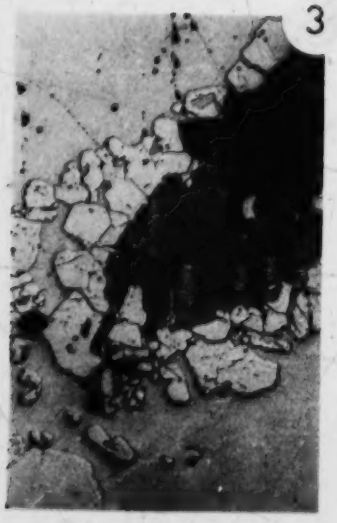
PLATE III

1

2

0·5 mm

3



1 mm

Plate IV

Cataclastic and Replacement Textures

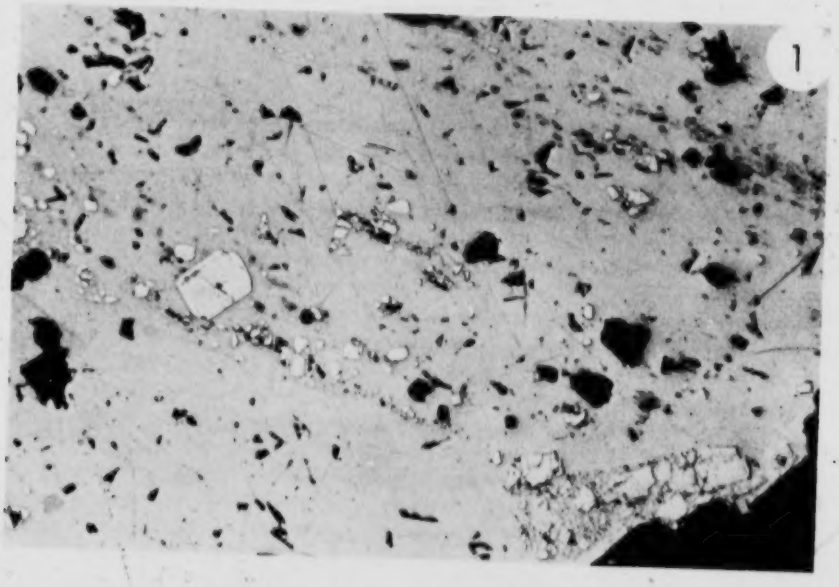
1) Euhedral pyrite fractured and alligned within the more ductile galena (sample SAVII from Scranton, 8x, reflected plane polarised light).

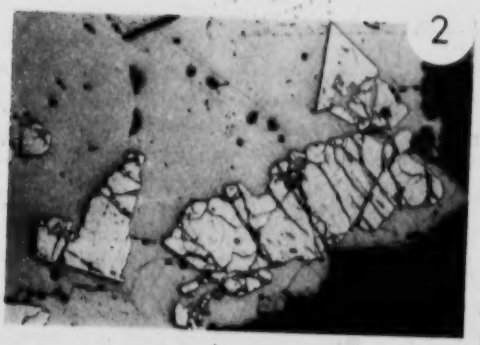
2) Fractured pyrite being replaced by galena. In this photograph one crystal of pyrite is in the octahedral form. (Sample 5702-SI from Scranton, 4x, reflected plane polarised light).

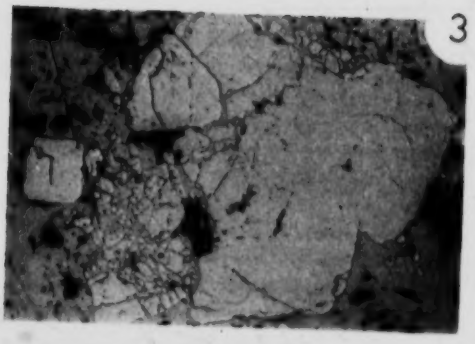
3) and 4) Distinctly euhedral pyrite cubes being selectively replaced by galena. (Samples SHI and 5702-SI from Scranton, 4x, reflected plane polarised light).

5) Replacement texture from Flint ore showing included blebs of sphalerite being replaced by pyrite. (Sample FAVI, 8x, reflected plane polarised light).

PLATE IV







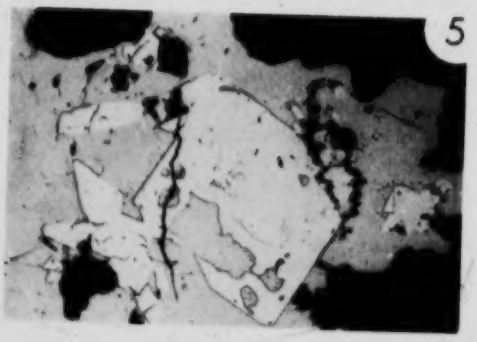


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Plate V

Vigilant Ore

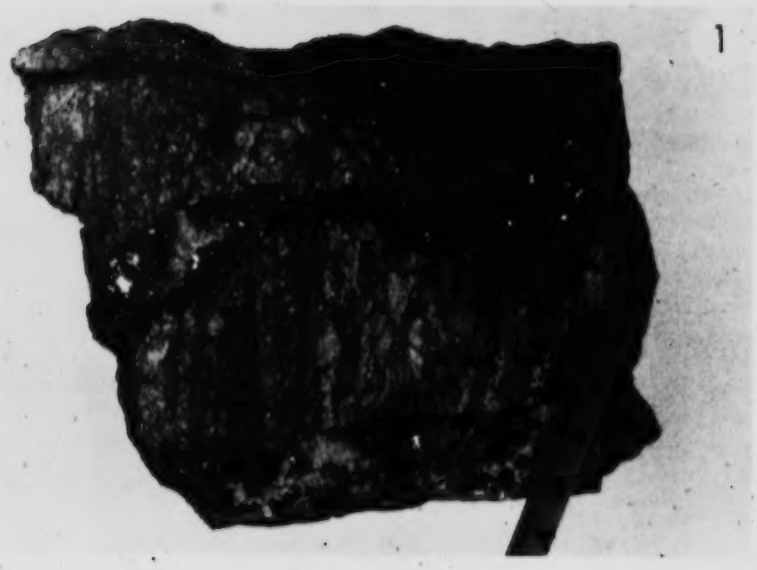
1) Hand specimen showing the ore from the Woman's Vein. Coarse-cubic galena and pyrrhotite form the dominant ore minerals within a calc-silicate host rock. Distinct replacement textures can be observed.

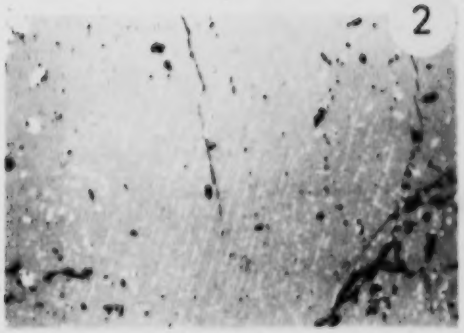
2) Exsolved chalcopyrite within sphalerite showing a distinct orientation (Sample WVH, 8x, reflected plane polarised light)

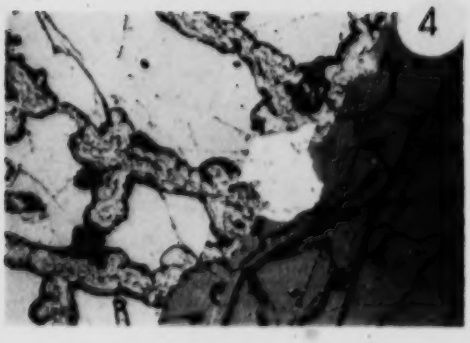
3) Distinctly euhedral quartz and arsenopyrite apparently growing at the expense of galena (Sample WVD, 8x, reflected plane polarised light).

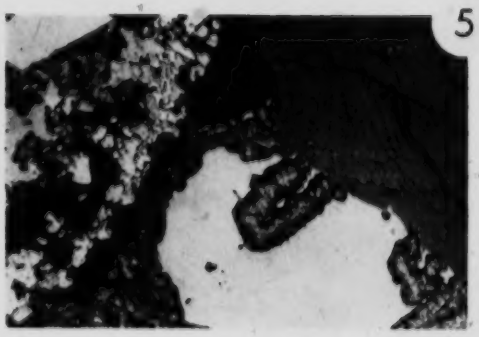
4) and 5) "Birds-eye" replacement textures formed by the alteration of pyrrhotite to a fine-grained intergrowth of pyrite and marcasite (Samples WVD and WVF, 4x, reflected plane polarised light).

PLATE V









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CHAPTER VI

MINERALOGY AND PARAGENESIS AT THE SCRANTON DEPOSIT

The Scranton deposit is typical of the "dry ores" found within fissure systems in the Nelson batholith. Little (1960 p. 123) includes the deposit in his lead-silver ore assemblage (plus zinc and minor gold). He notes that "mineral deposits containing lead and silver, in that order of value, are with few exceptions confined to a single belt extending from near New Denver southeastward through the Slocan mining camp to the Scranton mine". Scranton is situated approximately midway between the Ainsworth and the Slocan mining camps but tends to be more like the latter.

The deposit exhibits brecciated composite veins, typically common in the resistant Nelson granodiorite which under stress tends to fracture rather than flow, with little or no dissemination of ore minerals into the wall rocks. As a result a large proportion of gangue at Scranton is rock gouge associated with quartz and minor carbonate. Staining (Evamy 1963) reveals that the carbonates are mainly low-iron calcite and siderite with a minor amount of iron rich calcite and ankerite $Ca(MgFeMn)(CO_3)_2$. The quartz is usually white and massive but occasionally forms coarse crystalline aggregates in small vugs

indicating some minor open space crystallisation.

The major economic ore minerals are galena and sphalerite with lead, zinc, gold, silver and cadmium being extracted from the ore concentrate. The sample lead and zinc concentrate returns from the Scranton ore sent to Trail, B.C., give a good indication of the ore mineralogy. (Table 1).

Galena, the predominant ore mineral, occurs as both fine grained massive pods and also as coarse cubes. The sphalerite is generally massive and dark brown in colour. Massive and disseminated euhedral pyrite is very common throughout the full length of the fissure system. Minor arsenopyrite is present at the Pontiac and tetrahedrite, native silver and pyrargyrite have all been documented.

As no silver bearing minerals were observed in either hand specimen or on examination by reflected light microscopy the A.R.L. EMX microprobe, in the Geology Department at the University of Alberta, was used to determine the form of the silver-bearing phases. Using the wavelength dispersive mode the presence of silver was distinguished using the scanning electron beam and an audio-detector. Maps of the silver and gold concentrations were photographed using the oscilloscope camera (see Plate VI and VII). The silver was found to be mainly in discrete equant crystals (approximately 100 um x 30 um size) of silver

Plate VI

Electrum Within Scranton Ore

Sample 5702-SI, showing a highly reflective phase, was scanned for silver, gold and iron and a map of each of the concentrations was photographed using the oscilloscope camera attachment on the electron microprobe.

1) Silver scan, 10,000 counts x8 (ie. 6 u m per cm)

2) Gold Scan, 10,000 counts x8

3) Iron Scan, 50,000 counts x8

From the series of three photographs it can be seen that the gold occurs as an electrum. In this instance the electrum is included within pyrite.

PLATE VI

Plate VII

Argentite within Scranton Ore

This series of three photographs were again taken on the electron microprobe using the scanning oscilloscope. Here they show silver occuring as argentite (Ag_2S) adjacent to galena.

1) Lead Scan, 15,000 counts x 2 (ie. 24.2 um per cm)
2) Sulphur Scan 15,000 counts x 2
3) Silver Scan 9,000 counts x 2

PLATE VII

sulphide inferred to be argentite, Ag_2S, as only Ag and S were present. Argentite was also identified forming small inclusions within the sphalerite (less than 6 um x 3 um in size) and only a trace of silver was found in the galena. According to Mercer (1976) this is a common situation in lead-zinc Mississippi Valley Type deposits in contrast to stratiform massive sulphides where the silver, as expected, is strongly enriched in the galena (Hall and Heyl 1968). Taylor and Radtke (1969) have suggested that the silver in sphalerite is present as discrete micron size grains of silver sulphide, as suggested in this work, with silver contents up to 60oz/ton being documented. The concentrate returns from Scranton indicate a small but significant proportion of the silver is associated with the sphalerite. A minor amount of silver is also associated with the gold in the form of electrum which is found as both equant, irregular allotriomorphic crystal grains and in thin fractures within the other sulphides. The presence of electrum in fractures and its overall coarse grain size,(often up to 500um x 20 um) suggests remobilisation and recrystallisation (Ramdohr 1969). The silver content of the electrum varies approximately from 30-50% (ie. the fineness of gold is 500-700 typical of hydrothermal vein deposits - Bateman 1951)

and is a true electrum (Ramodohr 1969). All the gold appears to be in the form of electrum (see Plate VI) with none being found in the pyrite as originally proposed (Cairnes 1935).

Because of the extensive replacement of the sulphides, only an approximate paragenetic sequence has been established for the ore assemblage. Quartz and pyrite both appear to have a long depositional history being deposited throughout the paragenetic sequence. An early pyrite phase appears to act as the basis for the later main sulphides to be deposited by conditioning the site of precipitation. As a direct result replacement textures are very common. Early euhedral pyrite crystals show varied degrees of replacement by the later galena and to a lesser extent sphalerite. The order of replacement is, however, obscure with an obviously later pyrite phase replacing the galena and sphalerite often rimming the earlier more massive sulphides. The galena appears to be partially zoned as parts of some crystals are selectively replaced by pyrite. Pyrite replacement is also indicated by the presence of unreplaced minor blebs of galena, sphalerite and electrum within idiomorphic pyrite.

Abundant euhedral disseminated pyrite is found within the quartz and carbonate gangue and also forms late stage thin veinlets, associated with quartz, that crosscut all the earlier phases.

A very dominant feature of the Scranton ore association and one which obscures the paragenetic sequence is that of movement within the fissure system. Cataclasis, which in turn promotes further replacement, is a common phenomenon in the ores at Scranton. Pyrite, and to a much lesser extent sphalerite, deforms almost exclusively by brittle fracture (Atkinson 1975) with its strength depending strongly on pressure and only slightly dependent on the temperature. Galena, however, responds to stress by plastic deformation and is smeared along fractures within the broken euhedral pyrite crystals. At low temperatures and pressures ductile cataclasis of galena is common with intergranular shearing and local brecciation (Salmon, Clark and Kelly 1974). The formation of coarse cubic galena and occasionally vuggy quartz, which must have been developed in open cavities with little or no disturbance, suggests the pressures involved during mineralisation were also relatively low and probably somewhat localised. Brecciated vein matter recemented by later vein material

is characteristic of such composite veins and indicates that mineralisation probably took place at intervals resulting from minor deformative movements. Repeated intrusion of mineralising solutions is also indicated by the replacement textures observed.

The final solutions appear to have been carbonate-rich, depositing calcite in thin veinlets infilling porous fractures within the ores and nearby host rocks.

Associated deposits

i) The Flint Mine

The Flint mine property is situated at the head of Dago Creek a northwest tributary of Keen Creek on the eastern Flank of Mt. Carlyle. Like the Scranton deposit the Flint deposit is found in a composite shear zone close to the margin of the porphyritic Nelson granodiorite. The history of development of the mine is documented by Cairnes (1935). Banded ore structures are particularly common with siderite, quartz, cubic and fine-grained galena, sphalerite, pyrite and arsenopyrite. Sphalerite, the dominant sulphide, and galena both contain equant gold particles generally less than 20 um in size. Ore microscopy shows the sphalerite to contain dark brown internal reflections and also abundant exsolved chalcopyrite. Again pyrite and arsenopyrite, usually associated with massive white quartz, show replacement by and, to a lesser extent, of the more massive sulphides making the order of deposition diffi-

cult to estimate. However the sphalerite and galena appear to be earlier than at least some of the more iron rich sulphides. The ore shows numerous fractures, chlorite covered slickenslides and ore brecciation. Calcite (and ankerite?) veining appears to be the final stage of deposition partially infilling the earlier brecciated porosity.

ii) The Vigilant (Woman's Vein) System

These deposits are situated at the northernmost extremity of the Ainsworth mining camp, in the Lower Woodbury Creek area. Numerous small veins and fractures, generally striking east and dipping steeply southwards are to be found in the Early Bird Formation, of medium-grained calcareous and siliceous mica schist, of the Lardeau Group (Fyles 1967).

The "Woman's Vein" about 200 metres north of the Vigilant mine adit portal, sharply cross cuts a medium grained calcareous-silicate host rock. The vein, only recently explored, is about 10cm wide, has an indeterminate length, strikes north 73° east and is almost vertical. An earlier small quartz vein shows that a minor amount of displacement has occurred along the fissure with a right lateral offset of about half a metre. The presence of a thin gouge of crushed host rock, ankerite, calcite, and quartz with minor disseminated pyrite

indicates the displacement probably occurred during the mineralisation process.

Massive cube galena is the commonest sulphide with abundant pyrrhotite and pyrite and less commonly arsenopyrite, sphalerite and chalcopyrite. Gold in the form of equant, irregular fine-grained electrum and argentite are common. The sulphides show a very distinct contact with the carbonate rich host rock but some remobilisation of calcite has occurred. Minor calc-silicate "veining" can be seen within the marginal parts of the sulphides. Carbonate replacement appears to extend laterally away from the vein to a small degree with minor sulphide mineralisation, particularly pyrite, up to about 40cm. This suggests that the mineralisation is both fracture filling and replacement type. Both the sulphides and the gangue minerals are commonly in well formed crystals lining small vugs or in crustified textures.

The pyrrhotite is usually found at the margins of the vein and was probably early in the paragenetic sequence which is also suggested by the ore microscopy. As at the Flint and Scranton deposits the initial mineralisation is again iron rich. Following the pyrrhotite the main sulphide phase of galena, sphalerite and pyrite was deposited by selective solution of the

carbonate horizon. Chalcopyrite is extensively exsolved within the iron rich sphalerite. A slightly later phase of intrusion involves quartz veining with associated pyrite and arsenopyrite which crosscuts the earlier sulphides. These later sulphides are apparently responsible for extensive replacement of the earlier sulphides, again particularly the galena. Euhedral quartz has also apparently grown at the expense of the galena (see Plate V which also shows typical Vigilant ore). Following this quartz-rich phase, or associated with it, appears to have been movement along the fissure. Cataclastic pyrite and sphalerite fragments, caused by brittle fracture, occur within apparently undeformed galena which has flowed plastically. As at Scranton this cataclastic movement probably further enhanced the replacement effects.

The pyrrhotite has been subject to alteration along minor fractures producing a "birds-eye" texture (Ramdohr 1969 p. 532). This texture consists of both oval concentrically shelled bodies and simple alteration along fractures to a very fine grained marcasite and pyrite intergrowth (see Plate V). Minor goethite alteration of pyrite occurs along the fractures.

CHAPTER VII

ELEMENTAL GEOCHEMISTRY OF THE SULPHIDES

To test the amount of possible sulphide impurity within the sulphur isotope samples, these specimens were analysed using the Perkin Elmer 503 Atomic Absorption Spectrometer within the Geology Department at the University of Alberta. These analyses not only gave an indication of the "relative" purity but also an estimate of their elemental geochemistries. A few samples were also checked using the energy dispersive analysis method (E.D.A.) on the ARL EMX microprobe at the University of Alberta.

a) Scranton

The Sphalerite (see Table 2) has a low copper content and does not show any exsolved chalcopyrite (cf. Flint). As indicated by the dark brown colour and internal reflections, the sphalerite was found to have a high iron content (occasionally up to 5%). The higher lead values than expected are probably due to galena being present as an impurity in the sample. E.D.A. microprobe analysis shows trace quantites of cadmium within the sphalerite (but the concentrate returns indicate significant amounts are present). Except for the low copper values, the sphalerites from Scranton correspond reasonably well to

Table 2 Minor Element Geochemistry of Sphalerite (using A.A.S.)

	1a	1b	1c	2	3	4	5	6
Ag	100-1,000	100-1,000	100-1,000	100-300	10-50		Tr-12*	5*
Cd	4,000	4,000	4,000	1,000-3,000		600-3,000		
Co				50-300	10-100		11-22	25
Cr				20			Trace	Trace
Cu	1,000-5,000	1,000-5,000	5,000	10,000	very varied up to +1%		11-212	8,000
Fe	often +5,000	often +5,000	+5,000			9,000-49,700	20,000-52,000	42,400
Mn	500-1,000	Trace-100	Trace		very varied up to +1%	90-1,000	31-155	159
Ni						0-50	0	0
Pb	Trace	500-1,000	100-1,000	10-3,000 average 200			(3,000-11,000)	(8,000)
V	Trace						0	0

All values in ppm unless stated

Bracketed values indicate impurity

*Innaccurate results due to precipitation of silver chloride?

1. Warren and Thompson (1945)
   a) Slocan District - averaged values
   b) Ainsworth District - averaged values
   c) Cork Province, Keen Creek - averaged values
2. El Shazly et al (1947) - averaged values
3. Fleischer (1955) - commonest range of values
4. Evans, Campbell and Krouse (1968)
5. Scranton - this work - 3 samples
6. Flint - this work - 1 sample

those of Warren and Thompson (1954).

The galena samples were also analysed but because of sulphide impurities a meaningful interpretation of their geochemistry is not possible. However, a few general conclusions can be made. The silver content of the galena appears to be similar, if not slightly lower, than that of the sphalerite. These results however were suspect to "spikings" during the analysis and it is suggested that any silver present initially in either the galena or sphalerite was precipitated as silver chloride during the preparation of the sample. E.D.A. analysis indicates that only a trace of silver is present in the galena and that most of the silver is found within the sphalerite as inclusions of argentite and as discrete argentite and electrum grains within the ore.

Minor tungsten and a trace of tin have been reported from cut samples at the surface of the shear system. (British Columbia Department of Mines, 1954, p. A136).

b) The Flint Mine

The values obtained using the atomic absorption technique are documented in Table 2 for the sphalerite and Appendix 2 for the galena. The sphalerite is distinctly enriched in iron and copper due to the

presence of exsolved chalcopyrite. The values of cobalt and manganese are slightly higher than those of Scranton. The galena sample shows high zinc, copper, and iron values probably due to the presence of a minor amount of sphalerite.

c) The Vigilant Mine

Samples of pyrrhotite and galena were analysed as these were the major sulphides within the system. Little information can be extracted from the galena values except to note the exceptionally high manganese values (155 ppm) compared to Scranton and Flint. The pyrrhotite values (see Table 3) also show a relative enrichment in manganese. The high lead and zinc values reflect the level of impurity. The Co/Ni ratio of 5.28 suggests a late hydrothermal origin for the metals (Loftus-Hill 1967).

Table 3 Minor Element Geochemistry of Pyrrhotite

	1	2	3
Ag	X0	10	56
Co	40	Up to 5,000 common	66
Cr			49
Cu	100	100-1,000	11.5
Mn	X0	100-5,000	41.7
Mo		Very VariedUpto 1%	0
Ni	80	Very VariedUpto 1%	12.5
Pb	X,000		(10,366)
V			Trace
Zn			(1,650)

Values in ppm, unless specified

Brackets indicate impurity

1. Fryklund and Harner 1955
2. Fleischer M. (1955)
3. Women's Vein - pyrrhotite - one sample

CHAPTER VIII

SULPHUR ISOTOPE STUDIES

Sulphur has four stable isotopes with the following approximate abundances:

^{32}S	95.02%
^{33}S	0.75%
^{34}S	4.22%
^{36}S	0.02%

Of these the two most important isotopes are ^{32}S and ^{34}S, which differ in mass by about 6%, and normally sulphur isotope determinations measure the ratio of these two species.

The isotopic composition is expressed in terms of $\delta^{34}S$ which is defined as:

$$\delta^{34}S‰ = \left\{\frac{(^{34}S/^{32}S)\ \text{sample} - (^{34}S/^{32}S)\ \text{standard}}{(^{34}S/^{32}S)\ \text{standard}}\right\} \times 1000$$

The standard is the sulphur from the troilite in the Canon Diablo iron meteorite with $^{34}S/^{32}S = 0.0450045$ (Ault and Jensen 1962) which has been assigned a $\delta^{34}S$ value of 0.00‰ (ie. where $^{32}S/^{34}S = 22.22$). This has been found to be an appropriate standard because the isotopic composion of sulphur in mafic igneous rocks has been found to be very close to that of meteorites. Positive and negative $\delta^{34}S‰$ values

represent enrichment ("heavy") and depletion ("light") relative to the standard and consequently is a measure of the change in the isotopic composition since the introduction of the sulphur into the crust.

Variations in the isotopic composition of sulphur are formed due to fractionation between the sulphur source and the metallic sulphides precipitated. Three main factors determine this sulphur isotope fractionation; temperature, biologic interaction and solution chemistry.

a) Temperature

The isotopic fractionation between coexisting sulphur species in ore deposits is temperature dependent and therefore can be used as a geothermometer. Greater isotopic fractionation usually occurs at lower temperatures for sulphides and between sulphide-sulphate species.

The fractionation between coexisting sulphides was theoretically estimated by Sakai (1968) and Bachinski (1969) and based on chemical bond strengths indicated that ^{34}S enrichment should decrease in the order pyrite > sphalerite > chalcopyrite > galena. This fractionation trend has since been experimentally proven by Grootenboer and Schwarz (1969), Kajiwara and Krouse (1971) and Czamanske and Rye (1974) amongst others. Their results confirm that $\delta\,^{34}S$ values of

mineral pairs are directly related to the equilibration temperatures. The pyrite-galena pair is the most sensitive temperature indicator but unfortunately these two minerals are rarely found in equilibrium. Despite a smaller fractionation the sphalerite-galena pair usually gives the most reliable temperature estimates.

Particularly at low temperatures, there is a considerable variation in the estimated isotopic temperatures depending on the calibration used. Fluid inclusion data (eg. Rye 1974) appears to support the results of Czamanske and Rye (1974) with those of Kajiwara and Krouse (1971) being slightly too high and those of Grootenboer and Schwarz (1969) producing a temperature too low. More recent work by Smith et al (1977), however, appears to correspond favourably with that of Kajiwara and Krouse. The different temperature curves for the sphalerite-galena pair are shown in Figure 7.

b) Biologic Activity

An important cause of variation in the isotopic composition of sulphur in nature is the reduction of sulphate to sulphide catalysed through the metabolic activity of sulphate reducing bacteria (eg. the anaerobic Desulphovibrio desulphuricans). These bacteria excrete H_2S which is enriched in ^{32}S relative to the original sulphate. If this H_2S is then involved in sulphide

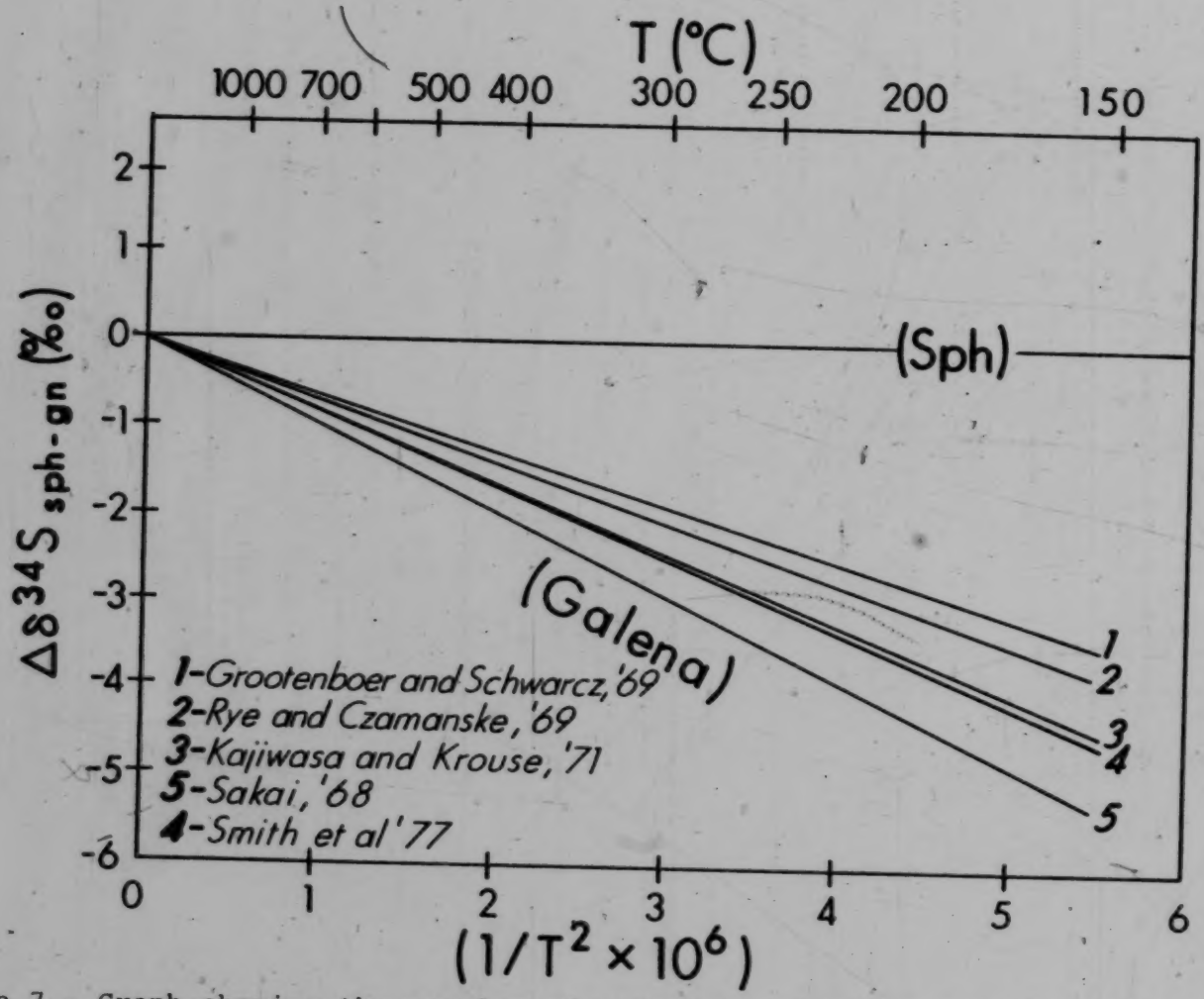


Figure 7 - Graph showing the results of calculations and experiments to determine the variation with temperature of sulphur isotope fractionation between galena and sphalerite.

precipitation then the sulphides will also be isotopically light. The extent of the fractionation, usually relatively large, is very variable and depends mainly on the rate controlling steps in the reactions by which the bacteria metabolise the sulphur. Biologic activity affecting the sulphur isotopic values is particularly prevalent in low temperature environments.

c) Solution Chemistry

Work by Sakai (1968) and Ohmoto (1972) has shown that the isotopic composition of sulphur in hydrothermal minerals is not purely a function of temperature and biological activity. This new data emphasised the importance of fO_2 and pH conditions, in altering the isotopic compositions, which in turn affects the proportion of the oxidised and reduced sulphur species in solution.

Large variations in the $\delta^{34}S$ values, previously interpreted as an indication of biogenic sulphur, could be explained by relatively slight variations in the $f O_2$ and/or pH of the ore forming fluids during ore deposition. Similarly near zero per mil values would not necessarily indicate a magmatic origin for the sulphur.

Rye and Ohmoto (1974) reviewed the overall state of sulphur isotope geochemistry and in particular emphasised the importance of the composition of the source sulphur and the solution chemistries in determining the isotopic abundances.

Previous Work Done on Sulphur Isotopes

A few sulphur isotopic values have been reported for the Bluebell Mine by Ohmoto (1971). He reports the isotopic composition of sulphur in sulphides (3 galena, 3 pyrite, 4 pyrrhotite and 14 sphalerite samples) show a gradual decrease in their $\delta^{34}S$ values from approximately + 11.5‰ for sphalerite from the early stage of Period II to around +5.5‰ for sphalerite from the later stages of Period III. Ohmoto notes that one pyrite sample from the very last stage has a $\delta^{34}S$‰ value of -23.6‰. The nature of the Bluebell hydrothermal fluids was briefly discussed by Ohmoto and Rye (1970). They suggest that the major aqueous sulphur species are H_2S and/or HS^-. Any changes in the pH and fO_2 environments of the ore fluids were apparently not reflected in the $\delta^{34}S$ variation (Rye and Ohmoto 1974) as the range of values is relatively narrow.

Ohmoto (1971) suggested that the sulphur isotope data indicates mixing of two different sources of sulphur, one with a $\delta^{34}S$ value of + 12‰ or greater and the other with a $\delta^{34}S$ value of -23‰ or lower. Rye and Ohmoto (1974) state that the sulphur in the Bluebell deposit was probably a mixture of sulphur from evaporites and sedimentary sulphides. The reasonably narrow range

of $\delta^{34}S_{\Sigma S}$ values for the Bluebell fluids indicates that the different sulphur components were well mixed in the hydrothermal solutions.

Present Work

A total of 30 sulphide specimens from the Vigilant vein system and the Flint, Scranton and Bluebell mines have been analysed and their sulphur isotope composition determined (see Figure 8). The overall spread in the $\delta^{34}S‰$ values is from -11.2‰ up to +8.2‰. However, very distinct trends are visible as can be seen in Table 4.

In all cases the sphalerite is isotopically heavier than the galena as predicted from their bond strengths (Bachinski 1969). This strongly suggests that sulphur isotope equilibrium has been approached by these minerals during their deposition.

Figure 9 plots the difference between the coexisting sphalerite and galena isotopic compositions ($\Delta^{34}S$) against their individual per mil values. By extrapolating the points plotted backwards towards the x axis the position of the intersection is an indication of the nature of equilibration. In this case there is a sharp intersect of each of the projected lines directly on the x axis, where $\Delta\delta^{34}S=0$, and indicates that the sulphide pairs are each in equilibrium

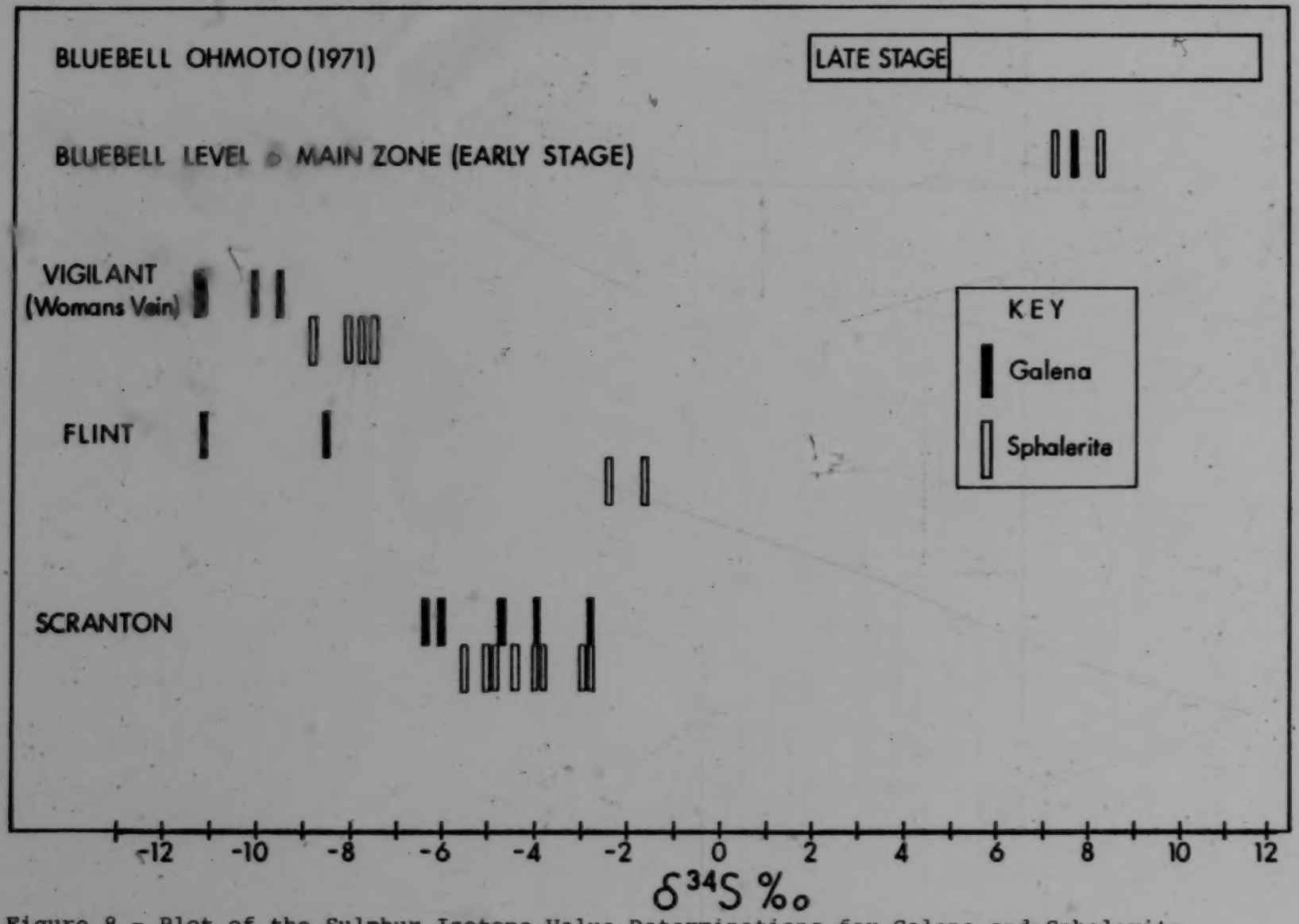


Figure 8 - Plot of the Sulphur Isotope Value Determinations for Galena and Sphalerite.

Table 4 Showing Variation in $\delta^{34}S$ Values

	Range of Values (‰)		Mean (‰)	
	Sphalerite	Galena	Sphalerite	Galena
Scranton	-5.5 to -2.8	-6.3 to -2.8	-4.2	-5.0
Vigilant	-8.8 to -7.6	-11.2 to -9.5	-8.1	-10.5
Flint	-2.4 to -1.5	-11.1 to -8.4	-2.0	-9.7
Bluebell	7.3 to 8.2	4.3 to 7.8	7.7	6.1

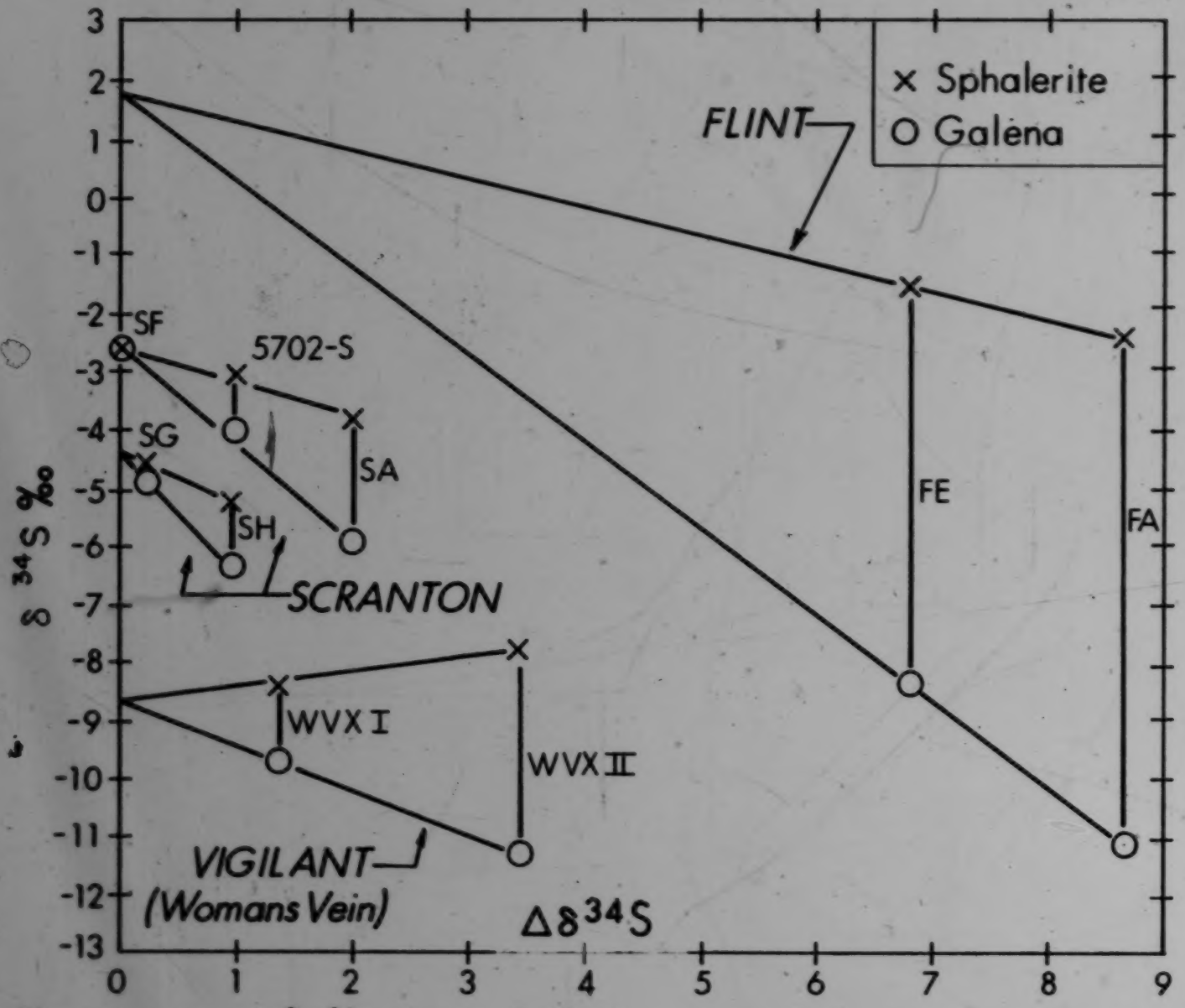


Figure 9 - Plot of $\delta\ ^{34}S_{\text{‰}}$ versus $\Delta\ \delta\ ^{34}S$ for sulphide pairs.

(Dr. H. R. Krouse and Dr. Grinenko - personal communication). This point of intersection also represents the initial sulphur isotopic value for that particular suite of sulphides. Again it can be seen that each deposit appears to have its own slightly different isotopic characteristics.

The relative shapes of the plots for the Flint and Scranton deposits are similar and suggests similar solution chemistries. The relatively steep slope shows that there was at least one other component in the system with a higher $\delta^{34}S$ value that is no longer represented. After plotting the values for the Scranton deposit it was concluded that two different"isotopic systems" were involved. It is suggested that there was a dual intrusion of hydrothermal solutions which precipitated the sulphides each of which appear to be in equilibrium. Each major impulse of mineralising solutions had a slightly different initial isotopic composition. The plot for the Vigilant system shows an initial $\delta^{34}S‰$ value of about -9‰, i.e. significantly lighter isotopically than the other deposits. A change in the fS_2, fO_2, pH etc., with the mean $\delta^{34}S$ value remaining the same, would still give the same intercept on the x axis. (Dr. H. Krouse- personal communication).

The few Bluebell analyses correspond well with those of Ohmoto (1971) and further confirm the large contrast in sulphur isotopic values with those of Scranton, Flint and Vigilant.

Temperature

Table 5 tabulates the $\Delta \delta^{34}S‰$ values and from these the temperature can be directly calculated using the experimental curves in Figure 7.

As can be seen in Table 5 the temperatures calculated using data from Kajiwara and Krouse (1971) are significantly higher, especially at lower temperatures, than those of Rye and Czamanske (1969) whose calibration was used in the present study.

Figure 9 indicates that the sphalerite-galena pairs used in the analyses appear to be in equilibrium. At high temperatures equilibrium is apparently nearly always attained (Ohmoto 1972). The very large $\Delta \delta^{34}S‰$ values for Flint however, therefore imply reequilibration of the sulphides at a much lower temperature.

A single galena-sphalerite pair from the Bluebell Mine (associated with early Period II knebelite from Level Six in the main Kootenay Chief ore zone) gives a $\Delta \delta^{34}S‰$ of 0.4‰ which indicates a temperature of 650°C. This high temperature corresponds favourably to earlier data published by Ohmoto (1971).

The Vigilant system gives a temperature of 160-390°C. These values show a large range in temperature but this can probably be attributed to the fact that

Table 5 Temperature Calculations from Galena-Sphalerite Pairs

Sample Name	$\delta^{34}S‰$	$\Delta^{34}S$	TempA°C	Temp B°C
WVXI Galena	-9.8			
WVXI Sphalerite	-8.4	1.4	450	390
WVXII Galena	-11.2			
WVXII Sphalerite	-7.7	3.5	240	160
FA Sphalerite	-2.4			
FA Galena	-11.1	8.7	< 100?	
FE Sphalerite	-1.5			
FE Galena	-8.4	6.9	< 100?	
SA Sphalerite	-3.9			
SA Galena	-6.0	2.1	360	300
SHI Sphalerite	-5.2			
SHI Galena	-6.3	1.1	580	500
SF Sphalerite	-2.8			
SF Galena	-2.8	0	650+	650+
SG Sphalerite	-4.6			
SG Galena	-4.9	0.3	650+	650+
5702-S Galena	-4.0			
5702-S Sphalerite	-3.0	1.0	600	510
BB Sphalerite	+7.8			
BB Galena	+8.2	0.4	650+	650

A - Using data from Kajiwara and Krouse, 1971

B - Using data from Rye and Czamanske, 1969

the vein is very small and the cooling effect of the host rock would be significant.

Temperatures for Scranton range from 300 - 510°C. The values for samples SF and SG are unreasonably high taking into account sulphide precipitation temperatures and wall rock alteration and therefore the values have been documented but not included in the temperature calculations (Dr. A. Sasaki personal communication). One possibility is that these particular pairs of minerals were not in equilibrium.

Conclusions From the Sulphur Isotope Analyses

Each deposit studied has unique sulphur isotope characteristics, with the Bluebell deposit being distinctly different isotopically from those of the Flint, Scranton and Vigilant deposits. The relatively narrow range for each of these latter deposits indicates that the sulphur components were well mixed in the hydrothermal solutions prior to deposition of the sulphides. Figure 9 shows, however, that the initial isotopic values of the mineralising solutions were different and indicates that the source areas were localised and at least partially separated.

The isotopic values for the Flint, Scranton and Vigilant deposits are all slightly negative and indicates that the majority of the sulphur was from a deep seated

igneous source (ie. released from silicate melts and from sulphides in igneous rocks). There is a need however to explain this negative deviation from a true magmatic source. Sasaki (1979? to be published in Contributions to Mineralogy and Petrology) states that sulphides derived from sulphur within mantle derived rocks are always a few per mil lighter isotopically. Jensen (1971) discusses the $\delta^{34}S$ values from metal deposits associated with the Cordilleran intrusives and states that negative values could be the result of isotopic equilibration between H_2S and SO_2 during the hydrothermal stage, with depletion of ^{32}S in SO_2 and enrichment of ^{32}S in H_2S. The former then oxidises, producing primary sulphates, which are lost from the system, while the latter is involved in the production of the metallic sulphides. Jensen (1971) shows that slightly negative $\delta^{34}S$ values are also seen in the sulphides from porphyry copper deposits associated with granodiorite intrusives and again indicates a sulphur source from the lower crust or mantle.

The $\delta^{34}S$ values for the Scranton, Flint and Vigilant deposits are very similar to those reported by Campbell et al, 1968, for the footwall of the Western Sector of the Sullivan deposit which shows a range of -10.4‰ to +4.7‰ with an average $\delta^{34}S$ of -2.2‰. They interpreted this footwall region as the conduit for the hydrothermal

solutions at the Sullivan deposit, starting from a deep seated magmatic source.

Mitchell and Krouse (1971) determined the isotopic composition of sulphur in galena from lead, zinc, fluorspar, barite mineralisation in the Yorkshire Pennines mostly in fault fissure deposits, epigenetic in style. They found the galena had a mean $\delta\,^{34}S$ of -2.2‰ and a very small range. These values again appear to be slightly enriched in ^{32}S relative to sulphides found in definate mantle origin, such as basalts, and also to the meteoritic mean. They concluded that this enrichment in ^{32}S indicates that sulphur from crustal rocks (eg. shales) might have been involved in the genesis of the ore-forming fluids.

It is indicated by the relatively negative values that there must be a "sedimentary" sulphur component in the ores from Scranton, Flint and Vigilant. The values from Vigilant, situated within the sedimentary sequence, as expected are more negative than for the deposits within the granite indicating a larger sedimentary component. Both the Scranton and Flint mines are near the sedimentary contact and therefore small convective hydrothermal systems could incorporate sedimentary sulphur into the fissure system.

Another possible explanation for the relatively negative sulphur values is to have a completely"igneous" sulphur source with "heavy" sulphide/sulphate deposition (eg. the Bluebell deposit) at a greater depth in the system than these deposits.

The temperatures indicated by sulphur isotopic calculations are generally high, but acceptable especially considering earlier data by Ohmoto (1971). The high temperatures further suggest an igneous hydrothermal environment for deposition of the sulphides.

It is suggested that temperature gradients within each of the deposits were relatively high but this could be an indication of slight disequilibrium at the time of precipitation or by later impulses of solution. Figure 8 indicates that the Scranton ores were probably deposited by two main periods of hydrothermal solutions of slightly differing isotopic composition. Multiple intrusion of mineralising solutions is also suggested by field observations.

CHAPTER IX

FLUID INCLUSION STUDIES

Description of the Fluid Inclusions

a) Scranton

Most of the inclusions observed within the quartz samples were extremely small (less than 30 um - see Plate IX with numerous secondaries along abundant fractures and shears.) Within the primary inclusions the vapour phase was small (1 to 2 volume %) with little variation. In larger inclusions (30 to 75 um) halite was common (1 to 2 volume %) with occasional sulphate crystals. Halite was identified by its cubic form, isotropism and its low thermal coefficient of solubility compared with sylvite. The sulphate crystals were typically elongate and distinctly anisotropic. The presence of CO_2 was suggested by a thick dark meniscus around the vapour phase caused by the relative differences in refractive indices. Secondary inclusions were abundant within the samples and were usually somewhat larger in size (up to 150 um). They had similar vapour volumes to the primaries but smaller halite crystals and an absence of sulphates suggested a lower salinity. Non-nucleation may account for the lack of daughter crystals which are often not seen within the smaller

primary and secondary inclusions. Very little necking appears to have occurred. Inclusions were described from the Scranton deposit but because of their small size were not suitable for experimentation and therefore samples from the old Pontiac Mine, about 300m to the northeast of the main Scranton workings and on the same mineralised shear, were utilised.

b) Flint

The inclusions observed from Flint, in both sphalerite and quartz samples, were significantly larger (up to 250 um) and more varied than those from Scranton. There is a large variation in the vapour phase from < 1 volume % to > 90 volume %. Two distinct groups, however, are discernible: the majority of the inclusions having vapour phases from 1 to 10 volume % (Type A) and another group essentially gas rich with approximately 70 volume % (Type B). The salinity of the inclusions also appears to be very varied. The Type B inclusions never contain halite crystals in direct contrast to Type A which may contain up to 50 volume % daughter crystals but usually about 5 volume %. Halite and sylvite (isotropic, cubic but with distinctly rounded crystal edges and a relatively high thermal coefficient of solubility) are both common. Sulphate, calcite (distinctly rhombohedral outline and

anisotropic and an opaque, non magnetic (oxide?) mineral are rarely present. Varied amounts of CO_2 were confirmed within the samples from Flint expecially associated with the more gas rich inclusions. A thin crescent shaped "fillet" (Roedder 1972) was observed within the vapour phase of a few inclusions which homogenised after heating with a hair-drier (ie. when heated above about 35°C). In a small number of inclusions the CO_2 phase is extremely large with a small vapour bubble, showing brownian movement, within the CO_2 liquid (see Plate VIII) There was little necking of the primary inclusions and, compared to Scranton, few secondary inclusions were present. The observed ratio of Types A to B is thought to be a poor indication of their proportion in the original ore forming fluid (Roedder 1967).

One sample from Flint (FB) was observed using the Scanning Electron Microscope to determine the nature of the internal structure of the fluid inclusions on a cleaved surface of sphalerite (see Plate X).

c) Vigilant

The inclusions from Vigilant are again slightly variable in their vapour phase and halite content but this is attributed to necking down which is common

Plate VIII

Fluid Inclusions from Flint

1) Primary inclusions showing a distinct variation in CO_2 concentrations. A CO_2-H_2O vapour phase (v), exhibiting brownian movement, occurs within CO_2 liquid (CO_2). The third phase in this inclusion is probably a low salinity brine.

2) Large primary inclusion showing sylvite (k), halite (H), and an unidentified opaque (0) phase.

3) Large primary inclusion containing sylvite (k), halite (H), sulphate (S), a dark gaseous phase (CO_2) and vapour (v). The abundance of daughter crystals, including some that are unidentified (?), suggests a complex saline brine.

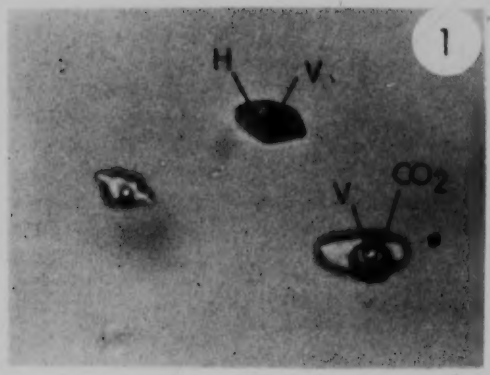
4) Photograph showing the large variation in the volume of the vapour phase. Type B is gas rich as compared to the more usual Type A with a smaller vapour phase. This variation indicates initial trapping of a heterogeneous solution.

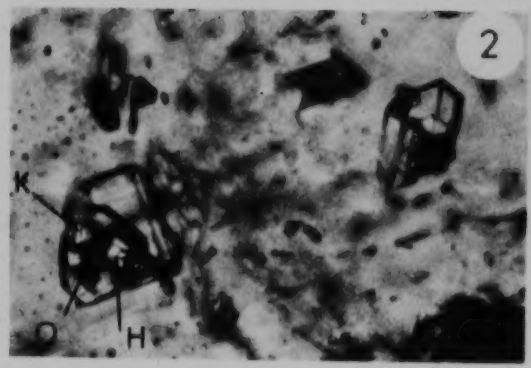
5) Negative (N) crystals indicate that this fluid inclusion occurs within quartz.

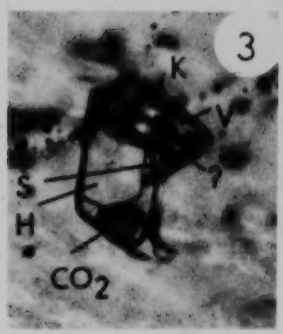
6) Inclusions again showing the heterogeneous nature of the mineralising solutions. Vapour (v), CO_2, opaque (o) and halite (H) can all be observed.

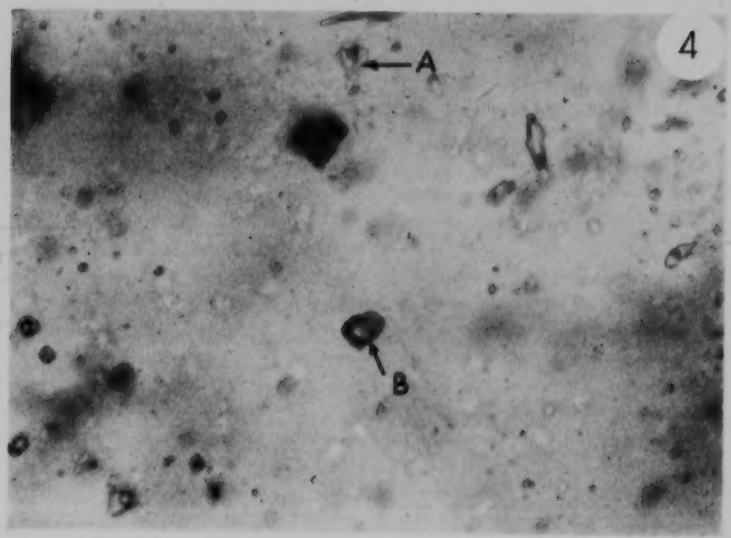
All photographs are 40x under plane polarised light.

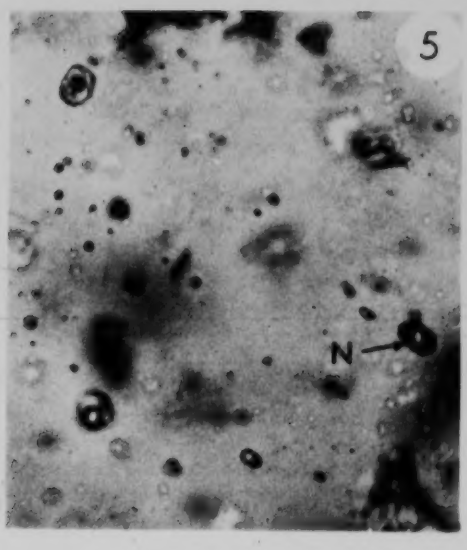
PLATE VIII











100 µm

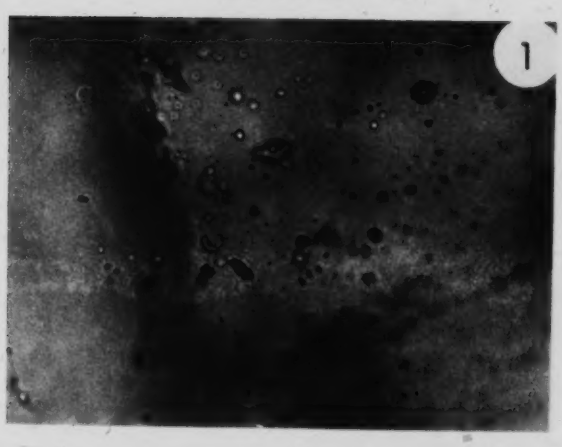
Plate IX

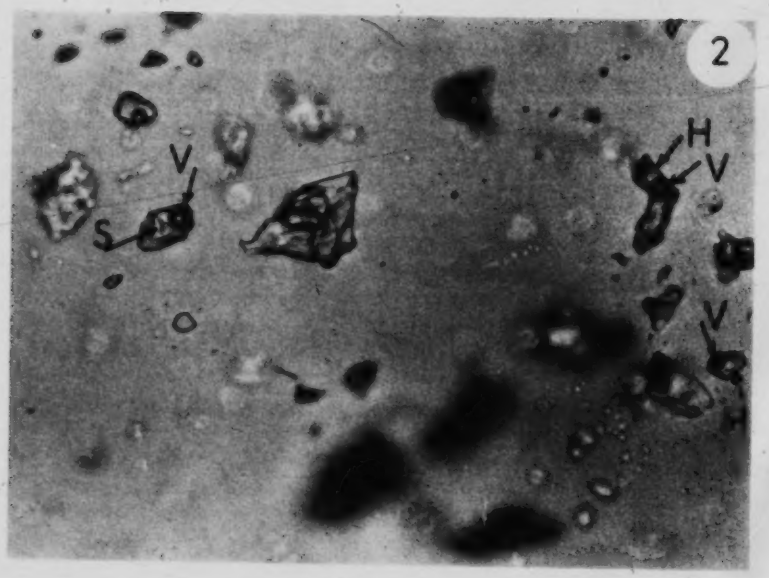
Fluid Inclusions from Scranton and Vigilant

1) Fluid inclusions within quartz from Scranton. Despite the poor resolution it can be observed that the inclusions are very small in size usually much less than 50 u m. (petrographic microscope, 40x, plane polarised light.)

2) Fluid inclusions within quartz from the Vigilant. Minor necking is indicated by the irregular shapes of the primary inclusions. Sulphate (s) and halite (H) daughter crystals infer the presence of a saline brine within the inclusions. (petrographic microscope, 40x, plane polarised light).

PLATE IX





100µm

Plate X

Scanning Electron Microscopy

Using the Cambridge 150 Scanning Electron Microscope within the Entomology Department, well cleaved surfaces of sphalerite were observed primarily to investigate the nature of the fluid inclusions in sample FB from Flint.

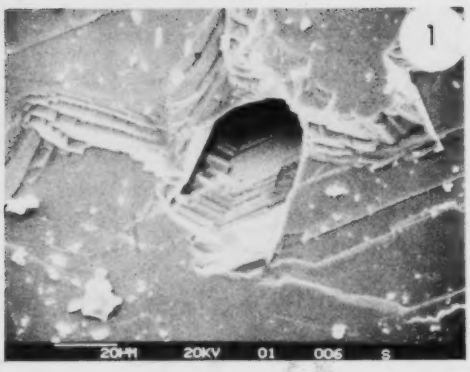
1) and 2) The internal three dimensional configuration of this particular primary fluid inclusion can be clearly seen. It shows growth striations within the inclusion forming a negative crystal shape. The fluid inclusions, initially formed under pressure, on fracturing would probably rapidly expel all its contents. This could account for the lack of daughter crystals within the remaining portion of the inclusion.

3) A primary fluid inclusion again showing faceted growth striations on its internal walls.

4) Associated with a large number of "empty" fluid inclusions was a botryoidal spongey precipitate. This might represent a residual chemical precipitate from the inclusions, caused by the sudden release of pressure.

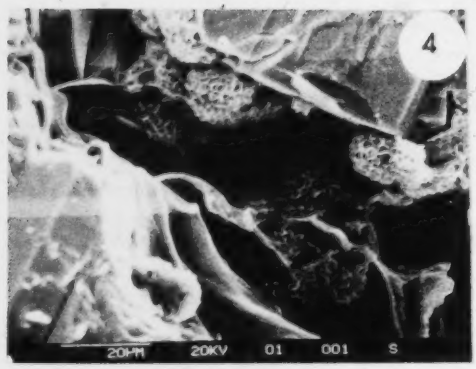
5) Minor carbonate crystals (distinctly rhombohedral) within an open space in the sphalerite.

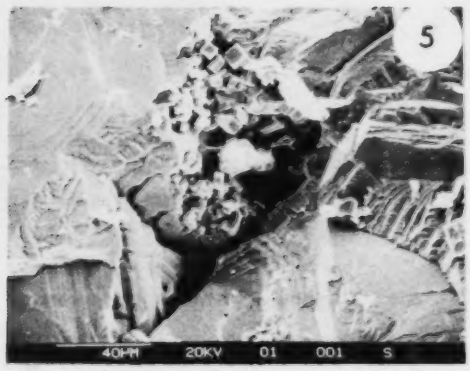
PLATE X


1
20µM 20KV 01 006 S

2
10µM 20KV 01 007 S

3
10µM 20KV 01 003 S


4
20µM 20KV 01 001 S


5
40µM 20KV 01 001 S

within these specimens. They vary in size from 30 um to about 150 um. The volume of the vapour varies between 4 to 6% with halite crystals generally small (less than 0.5 volume % on average). Although these solutions are definitely slightly less saline than those at Scranton the lack of halite crystals in the smaller inclusions may again reflect the failure of this phase to nucleate. No CO_2 was observed during the microscopic identification. A photograph of fluid inclusions from Vigilant can be seen in Plate IX.

Results of Crushing Experiments

Using a simple crushing device (see Appendix 4) it was found that high pressure gases were released into a glycerin medium from the fluid inclusions from Scranton and Flint. Very minor gas production was noted from within inclusions from the Vigilant. To determine the nature of the released high pressure gas, fragments of each specimen containing inclusions were crushed in a barium hydroxide aqueous solution medium. A white precipitate of barium carbonate was formed suggesting minor CO_2 is present in both the Flint and Scranton inclusions. No odor of H_2S was released on crushing which, if present, would be detected by the human nose which is an exceedingly sensitive detector being able to smell quantities of H_2S less than 10^{-10}gm. (Roedder 1971).

Determination of Organics

Rosasco, Roedder and Simmons (1975) have documented the successful application of laser-excited Raman spectroscopy to the identification and partial analysis of solid, liquid and gas phases in fluid inclusions. A similar application using the Raman spectrometer in the Chemistry Department at the University of Alberta was attempted using fluid inclusion samples from Scranton as well as from the Terra Mine, N.W.T., where organic liquids had previously been documented (Brame, 1977, Geology 607 project, University of Alberta). The results, however, were not encouraging because the apparatus did not respond to the very low organic concentrations within the inclusions.

Using the Nicolet 7199 Infra Red apparatus, also in the Chemistry Department at the University of Alberta, two definite carbon peaks were found to be present in the samples from Terra at 2920cm^{-1} and 2850cm^{-1}. The former value is indicative of methane and the other is a carbon-carbon bond structure of unknown size. Using the Terra sample as a standard no carbon peaks were isolated from the Scranton samples indicating that no organics are present. The expansion of this method related to the study of fluid inclusions would prove very fruitful especially as the apparatus can be put

under vacuum thereby also enabling the determination of OH and CO_2.

Salinity Estimates

As halite crystals are present, freezing experiments are not applicable and the brines must be greater than 26.3 NaCl equivalent weight % (Roedder 1967). It is therefore necessary to use a volumetric analysis method as described in the appendices.

The salinity of the Scranton brines, assuming that the volume % of each of the vapour and the halite is 2, is calculated to be 37 NaCl equivalent weight %. Assuming a volume of 5% vapour and 0.5% halite the Vigilant solutions have a salinity of 33 NaCl equivalent weight %. It is impossible to calculate the salinity of the solutions from Flint accurately because of the differences in the phases present. The initial solutions prior to boiling, however, were probably very similar to those of Scranton.

Another method that could be used to determine the salinity is the temperature at which the halite crystals dissolve; this can then be referred to solubility curves. It was found, however, that the halite crystals remained undissolved after the homogenisation of the inclusions at elevated temperatures. The delayed

dissolving of halite has commonly been reported and is probably caused by heating the specimen too rapidly causing disequilibrium or is due to the effects of post trapping changes (eg. hydrogen diffusion). As a result this alternative method was not used in the determination of the salinity.

Heating Experiments

By heating the fluid inclusions until the vapour phase homogenises with the liquid phase an estimate of the temperature of the mineralising solutions can be obtained, after allowing for a pressure correction. Because of the high salinity of the brines involved in the study, the observed volume of the vapour phase at room temperature cannot be used as an accurate estimate of the temperature of homogenisation. This is due to the compressibility of the saline fluids reducing the volume of the vapour.

a) Scranton

All the primary and secondary fluid inclusions homogenise in the vapour phase. Because of the small size of the inclusions it was fortunate that prior to homogenisation the vapour bubble usually exhibited brownian movement and therefore an accurate temperature could be obtained. The filling temperatures showed a

large range from 199.3 to 276.5°C (30 values) with the secondaries homogenising in the same range as the primaries, although many appeared to homogenise at the higher temperature range.

b) Vigilant

All primary fluid inclusions again show homogenisation in the vapour phase. The filling temperature recorded for 23 inclusions range from 174.9 to 223.1°C with a concentration between 190 and 210°C. Although inclusions showing necking were avoided in this study, this range is probably caused by it.

c) Flint

Despite the very large variation in the vapour phase, heating experiments were performed on these samples. For Type B (gas rich) inclusions homogenisation was always in the gas phase compared with that of Type A in the more usual vapour phase. The homogenisation temperatures range from 156.6 to 223.3°C (34 values) with a marked concentration between 200 and 218°C. Both Type A and Type B inclusions homogenised respectively by contraction and expansion of the gas phase over the full range of temperatures. It is suggested that this entrapment of coexisting liquid and gas rich inclusions results from heterogeneous boiling or effervescing. The variation in homogenisation temperatures

is thought to be due to a variation in pressure at the time of trapping, and not to large changes in solution temperatures (Roedder 1971). As no pressure correction is required for a boiling solution, the homogenisation temperatures reported are thought to be a good estimate for the true temperature of the hydrothermal mineralising solutions at Flint. Boiling would occur with a sudden pressure drop and when the vapour pressure became greater than the hydrostatic pressure. Growing crystals would become coated with gas bubbles shielding small areas of the crystal face from the liquid phase, thereby inhibiting growth, resulting in engulfing of gaseous inclusions. As a result two discrete inclusion types are found with no intermediate type present. Individual inclusions that trapped both liquid and gas, would give maximum depositional temperatures (Kelly and Turneaure 1970). Any CO_2 initially present within the salt solutions would be enriched in the vapour phase, as found, by boiling solutions.

Depth of deposition at Flint

As the hydrothermal solutions at Flint were boiling, it is possible, from knowledge of both the salinity and the homogenisation temperatures from fluid inclusions, to determine the depth at which this boiling occurred. The process utilises the boiling point

curves documented by Haas (1971) who calculated the temperature-depth relationships for an open vein system where an $NaCl-H_2O$ liquid is at the boiling point for the confining hydrostatic pressure.

Assuming that the salinity of the solutions at Flint was similar to the salinity of the solutions at Scranton (ie. approximately 35 NaCl equivalent weight %), then an extrapolation of the boiling point curves provided by Haas (1971) is required for these slightly higher salinities. With a temperature of 200 to 220°C, a depth of 200 to 310 metres is estimated. This depth calculation, however, implies a full hydrostatic column of dense brine and as nearer surface waters are likely to dilute the column, this estimate will be a minimum value. The vapour pressure in complex brines is significantly lower than in simple $NaCl-H_2O$ brines and therefore these depth estimates are again conservative. A depth of 400 to 500m is therefore estimated.

Temperature Determinations

As previously stated it is necessary to apply a pressure correction to the homogenisation temperatures when non-boiling conditions are present, in order to give an estimate of the true temperature of deposition.

Lemmlein and Klevstov (1961) and Potter (1977) have produced graphs of the pressure effects on various NaCl-H_2O solutions at different temperatures. Although graphs are not available for salinities above 30 NaCl equivalent weight %, at this salinity range the errors in pressure correction are very minor (Roedder 1971). This is because the capacity for expansion of the brines decreases considerably at high salt concentrations.

a) Scranton

By direct analogy with Flint, these solutions have not boiled, implying a depth of deposition greater than 500m with lithostatic pressure conditions (ie. approximately 2.6gm/cm^3). The inclusions (with homogenisation temperatures ranging from 200 to 275°C) would therefore have a minimum pressure correction of 180°C to 190°C to give a true temperature of 380°C to 465°C. Figure 10, adapted slightly from Potter (1977), shows the temperature corrections necessary, for a saline brine of about 25 to 35 NaCl equivalent weight %, as a function of the homogenisation temperature and pressure.

b) Vigilant

Again using the depth constraints provided by Flint, a true temperature of deposition can be estimated.

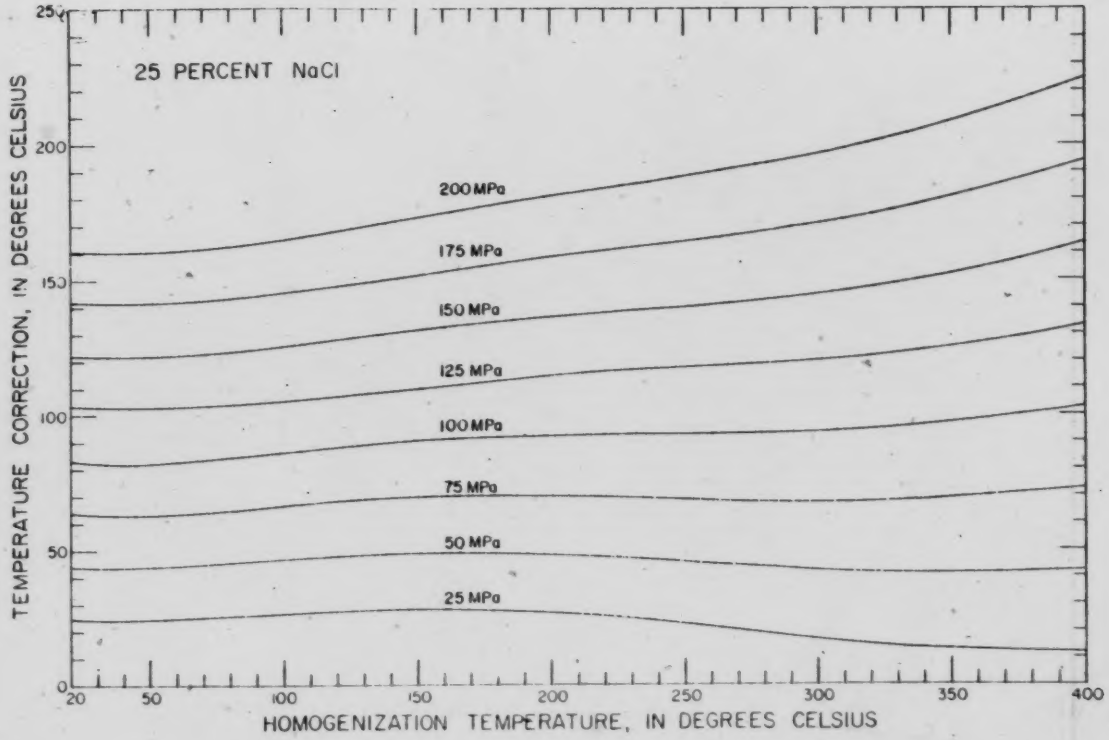


Figure 10 - Temperature correction for a 25 percent or greater NaCl solution as a function of homogenisation temperature and pressure (from Potter 1977).

The pressure correction for a 33 NaCl equivalent weight % brine at 190 to 210°C (uncorrected) would be 180°C to give a temperature estimate of 370 to 390°C.

Conclusions

The mineralising solutions at the Scranton, Flint and Vigilant deposits were all saline brines with a density greater than 1.0gm per cm^3. The presence of halite and, more rarely, sylvite, calcite and sulphate indicate NaCl rich brines, with K^+, Ca^{2+}, CO_3^{2-}, and SO_4^{2-} ions also present in smaller amounts. The salinity of the hydrothermal solutions at each deposit was very high, namely about 35 NaCl equivalent weight %. This suggests that the Vigilant and Scranton solutions were also probably quite close to the boiling curve, as suggested for the solutions at Flint. Very high salinities are possibly indicative of a magmatic source (Roedder 1971). Ohmoto and Rye (1970) and Ohmoto (1971) have reported that salinity values for the Bluebell Mine, gradually reduce from 15 weight % at 550°C to 3 weight % at 320°C.

Takenouchi and Kennedy (1965) state that if solutions contain a considerable amount of dissolved salt it would dissociate into two immiscible fluids

one rich in CO_2 and the other in water. The solubility of CO_2 decreases rapidly as the amount of salt increases and as a direct result the fluid inclusions would only contain minor CO_2. In a boiling fluid however, any CO_2 would be concentrated in the gaseous phase, as was found at Flint.

The primary and secondary inclusions at Scranton indicate a pressure-corrected depositional temperature of 380 to 465°C. The presence of secondary inclusions with a similar temperature to that of the primaries indicates that during deposition there was abundant shearing caused by tectonic movements. The secondaries appear to have a slightly lower salinity and perhaps a slightly higher temperature of deposition - further indicating the possibility of the multiple intrusion of slightly different solutions. These estimated temperatures are slightly lower than the sulphur isotope geothermetric calculations (300 - 510°C).

At the Vigilant, there is considerable necking of the fluid inclusions. Nevertheless, a temperature of 370 to 390°C is estimated which coincides well with the upper value provided by the sulphur isotope determinations (390°C).

It is suggested that the Flint solutions were near boiling at the time of deposition, from the

evidence of variable heterogeneous trapping of a gas-rich phase and a salt-rich aqueous phase. The homogenisation temperatures are varied as would be expected from a boiling environment but there is a marked concentration of values between 200 and 218°C. No pressure corrections are required and a depth of deposition of 400 to 500m is suggested. Variations in the pressure conditions at the time of deposition could account for the boiling. If there was an alternation between lithostatic and hydrostatic pressure it implies that the hydrothermal passageways were occasionally blocked, with tectonic movements opening new channelways and releasing pressure. This sudden pressure drop would encourage boiling to occur. The first effects of boiling would be to precipitate the least soluble constituents such as the sulphides with more boiling producing the precipitation of high quantities of NaCl and other salts (Ridge 1974).

Numerous other examples of boiling have been documented particularly in porphyry environments eg. Roedder (1971), Moore and Nash (1974), Nash (1976). Vein deposits at some stage in their development also appear in many cases to have been close to the boiling curve. Rye and Sawkins (1974) have documented examples at Casapalca and Kelly and Turneaure (1970) in Bolivia.

CHAPTER X

ORIGIN OF THE MINERAL DEPOSITS

Early models for the origin of the Ainsworth and Slocan deposits, postulated that the hydrothermal ore solutions were directly related to the formation of the Nelson Batholith. Earlier work by Uglow (1917) was expanded by Bateman (1925), in a paper too often ignored by later workers, and then by Cairnes (1934), all of whom discuss the relationship of the deposits to their respective host rocks. Bateman (1925) states "if the dykes and sills have a genetic connection with the batholith then so do the veins" and Schofield (1920) the "ore bearing solutions emanating from the Nelson granite in the pneumatolitic stage of cooling." During intrusion of the batholith Bateman envisaged numerous cupolas extending into the folded roof sediments, producing fracturing of these overlying rocks resulting in the injection of porphyry dykes and then further fracturing and expulsion, from the unconsolidated interior of the batholith, of the metallising solutions.

Showing a response to the distance the solutions travelled a zonal distribution of the ore was postulated:

a) within the batholith the ores tend to be highly siliceous, the total amount of metallic minerals is generally less than the amount of gangue and the silver values are high. These ores typically form within composite breccia-veins.

b) within the roof sediments (eg. Slocan type), the ore-gangue is generally less siliceous with carbonate (mainly siderite and calcite) predominating. Corresponding to an increase in galena and sphalerite quantities is a decrease in the silver contents. These ores typically form within shear-veins.

As well as a zonal distribution a variation with depth was recorded with deposits commonly being more lead-and silver-rich in the upper portions grading downwards through a more zinc rich zone into a lowermost iron zone. This rapid change in mineral composition was taken to indicate steep thermal gradients with temperature being a strong controlling factor in mineralisation.

The zonal distributions were attributed to a mineralised zone, up to 610m thick (Cairnes 1934) related to the presumably highly variable batholith surface and bearing no relationship to the present day topography.

Hedley (1955) continuing this earlier trend stated that the events which lead to the formation of the ore are related to the emplacement of the Nelson Batholith. Hedley (1952), however, found that within individual lodes,although a vertical mineral zonation had been recognised,the same mineral sequence could also be found horizontally. This suggests that the deposits do not form purely as a result of the temperature variation but that other related factors, such as the structural control, are also extremely important in ore deposition.

Rice (1944) showed that the "dark coloured dykes and sills" within the Ainsworth camp are distinctly younger than the main phase of the batholith yet older, in most cases, than the mineral deposits. He indicated that the ore deposits "... cannot have been formed before Tertiary time. The evidence, however is not considered conclusive." Realising the complexities involved within the area Fyles (1966) stated that the "interplay of time, structure and physical-chemical conditions have resulted in a variety of geological settings." Fyles (1967) expanded his views in "Geology of the Ainsworth-Kaslo Area" noting again that the transgressive deposits crosscut a series of lamprophyre

dykes which are not deformed by the Laramide (Rocky Mountain) Orogeny in early Tertiary times (Ridge 1972). Hedley (1952) in ores from the Sandon area also notes that "all dykes are pre mineral." From this Fyles (1967) concluded that the transgressive deposits are younger than all facies of the Nelson and therefore are not related genetically. In an excellent summary he states"... there is no evidence of regional zoning related to the batholith. If ore forming solutions were derived from some magma chamber it must have been deeper than the presently exposed granitic rocks."

In all cases the age of the mineralisation has been related to field evidence involving crosscutting relationships with lamprophyre dykes. Very little work, however, has been documented on the relationships and age of these dykes. Nguyen et al (1968) have provided the only age available,indicating 169±6 Ma which, within the limits of their experimental precision, is the same as the age obtained for the Nelson Batholith. Most of the sediments within the central Kootenay Arc show ages, by K/Ar of about 50 to 60Ma. This has been attributed to reheating, resulting in argon loss, caused by the presence of plutons at depth (Dr. H. Baadsgaard - personal communication). Similar plutons, at a lower erosional level in the Kootenay Arc, are exposed in northeastern Washington and northern Idaho (Miller and Engels

1975 and Armstrong, Taubeneck and Hales 1977). One such pluton may have caused the high grade "regional" metamorphism in the Riondel area.

A regional variation of selected minor elements in sulphides, from the Slocan Mining Camp, has been shown to exist by Sinclair (1967) and Sinclair and Mathews (1969). Using trend analyses and residual maps they showed that regional patterns of silver in galena, tin in sphalerite and also Co/Ni ratios, arsenic, chromium and tin in pyrite were all similar. These values form coincident and superimposed "highs" ("basin" shaped for the Co/Ni values) situated in a northwest trend along the position of Hedley's Slocan Fold suggesting the structure and folding has played an important role in the localisation of the deposits. The values also suggest a central zone of high temperature, centred near Sandon, about three miles from the nearest Nelson outcrop at the surface. Sinclair (1967) indicates that this would make it difficult to relate the plutonic rocks and the mineral deposits to a single common magmatic source.

Analysis of the Slocan sediments by atomic absorption shows that lead, zinc, and copper are all at usual or only slightly above crustal levels. This suggests that the sediments were probably not the source of the metals involved in mineralisation (Cox, 1979).

The documented rock and ore lead isotope studies provide us with valuable information as to the origin of the mineralising solutions. Sinclair (1966) indicates that there is no correlation between lead isotopic composition and the minor element content of galena, age of wall rock, type of wall rock and geological type of deposit (ie. replacement or open space filling). Lead within each ore deposit has a uniform isotopic composition but there is often a large variation between separate deposits suggesting only local homogenisation and incomplete mixing of the solutions. Ore lead from the Slocan and Ainsworth camps is distinctly different from those deposits, apparently unrelated to the Nelson Batholith further to the north and the south (eg. the Sullivan and Salmo types). Data from the Bluebell Mine have been documented by Kanasewich (1962, 1968) which suggests that this deposit also belongs to the group related to the Nelson Batholith. All these ore leads are distinctly anomalous, having a multistage history with the radiogenic component being evolved in a system that closed at about 1500Ma ago (Sinclair 1966). Cox (1979), confirming this previous work, has added another interpretation to the data by recognising two distinct lead isotope populations.

One group which contains the Ainsworth and Bluebell deposits shows evidence of having lead derived from the Precambrian continental basement. Ore lead from the Sandon camp and deposits within the Nelson Batholith form a second discrete group with lead apparently derived from an oceanic source.

Margaritz and Taylor (1977) using hydrogen and oxygen isotopes have shown that the Nelson Batholith has undergone very large scale water-rock interaction. Their interpretations suggest, however, that this alteration was associated with the younger Eocene plutons. Any alteration related to Jurassic meteoric water circulation resulting from the cooling of the Nelson Batholith, if present, has been over-printed by this younger event. Taylor (1977) has shown that the size of a "hydrothermal cell" is usually somewhat larger than that of the pluton. By analogy one would therefore expect the very large Nelson Batholith to show primary water-rock alteration of Jurassic age. The situation is obviously only poorly understood especially as Margaritz and Taylor (1977) apparently only sampled marginal phases of the Batholith. With more detailed sampling and analysis of rocks from towards the centre of the batholith and also of ore material a better understanding could perhaps be arrived at.

The sulphur isotope results from this study suggest that the sulphur is igneous in origin with a small sedimentary component. This further enhances the possibility of the presence of a large hydrothermal cell (greater than 2000 square km?), related to the cooling of the Nelson Batholith, involving both the batholith and the surrounding sediments. Each deposit, however, shows unique sulphur and lead isotope characteristics which suggests that superimposed within this large hydrothermal system are small discrete mineralising sources. If the Nelson Batholith extends northwards under the Slocan Sediments, as is indicated by the large number of granitic intrusions, then a discrete "hot spot", centred near Sandon (Sinclair and Mathews 1969), can be accommodated within this interpretation in the form of a local hydrothermal cell.

The mineralising solutions were hot (up to 510°C at Scranton) and were made up of complex saline brines up to about 40 NaCl equivalent weight %. The deposits were precipitated close to the surface as indicated by the presence of boiling at Flint (which could account for the apparent re-equilibration of the sulphur isotopes at low temperatures). The presence of boiling can be directly related to the cooling and resultant crystallisation of the granitic

silicate melt. On cooling the water vapour pressure rises until the saturation isobar is attained at which stage boiling occurs (the "second boiling point phenomena" described by Mueller and Saxena 1977 p. 272). Further cooling then follows this isobar and as boiling continues, with only a slight temperature decrease, the residual liquid becomes slightly enriched in water until crystallisation is complete. The water content, however, in absolute amounts is reduced to about 0.5% in the final rock with about 1 to 5 weight % of water being released during crystallisation. Holland (1972) has shown that the classical ideas on the derivation of postmagmatic hydrothermal deposits are basically correct. The hydrothermal solutions produced during crystallisation contain a sufficient concentration of ore metals to give rise to the "observed base metal deposits" especially when the chloride contents of the solutions are high. A decrease in temperature and pressure along with the reaction of these solutions with the wall rocks would lead to the precipitation of the ore minerals.

It can be envisaged within the ore deposits in the Nelson Batholith that when the fissure systems, acting as conduits for the mineralising solutions, become choked, lithostatic pressures would be approached in the system and therefore boiling would be inhibited.

The hot solutions were probably injected into the fissures at high velocities, depending on the varying pressure conditions produced by simultaneous tectonic movement. The wall rocks, as a result, are only mineralised to a very limited extent with minor wall rock alteration. Fumarolic and hot spring emission, with sulphate precipitation, would probably have occured.

This study suggests that the Slocan, Ainsworth and Bluebell deposits are all related to the Nelson Batholith with the mineralisation, therefore, having an age of late Middle Jurassic. The Ainsworth and Bluebell deposits within the sediments adjacent to the Nelson Batholith, are further from the heat source and are therefore somewhat modified. In conclusion, these deposits show a very distinct contrast to other lead-zinc deposits within Western Canada.

Having a high temperature, magmatic origin these deposits are markedly different from Mississippi Valley type deposits. The best comparison can be drawn with the Sullivan deposit which is volcanogenic-exhalative in origin, with a relatively high temperature origin and a magmatic derived sulphur component, but which occurs in a sedimentary sequence of middle Proterozoic age.

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i) Major Elements (%)

	1		2	3	4		5	6		7	
	a	b			a	b		a	b	a	b
SiO_2	62.88	62.69	52.90	62.47	70.71	70.68	67.30	66.21	66.59	63.62	63.47
Al_2O_3	16.33	16.54	14.34	15.25	15.60	15.86	16.27	16.46	16.14	15.96	16.22
Fe_2O_3	5.39	5.39	9.37	6.01	1.69	1.68	3.03	3.35	3.36	5.12	5.23
MgO	2.03	1.96	6.32	2.46	0.38	0.38	0.68	0.76	0.74	1.83	1.80
CaO	4.84	4.91	7.49	5.24	1.62	1.59	3.08	3.54	3.50	5.18	5.21
Na_2O	4.14	3.87	2.78	3.78	4.34	4.20	4.60	4.72	4.66	4.66	4.49
K_2O	3.28	3.36	4.62	3.44	5.28	5.25	4.32	4.32	4.39	2.43	2.47
TiO_2	0.75	0.75	1.38	0.80	0.22	0.22	0.38	0.40	0.40	0.71	0.69
S	0.07	0.07	0.07	0.14	0.03	0.03	0.06	0.06	0.06	0.00	0.06
MnO	0.07	0.18	0.08	0.09	0.07	0.07	0.017	0.05	0.05	0.12	0.12
P_2O_5	0.23	0.26	0.66	0.33	0.07	0.05	0.10	0.11	0.10	0.29	0.25

a and b represent repeats of the same sample (splits of one powder).

All analyses were performed by Dr. J. G. Holland, University of Durham, England.

The key to specimen numbers can be seen in Appendix 1 iii.

ii) Minor Elements (ppm)

	1		2	3	4		5	6		7	
	a	b			a	b		a	b	a	b
Ba	1,261	1,224	1,175	1,262	1,264	1,348	1,174	1,192	1,180	841	868
Nb	39	39	24	25	11	12	22	24	25	36	35
Zr	265	250	172	192	114	123	197	223	224	268	260
Y	24	27	26	21	10	11	14	17	19	23	20
Sr	1,046	1,056	827	940	1,070	1,031	1,186	1,250	1,273	1,076	1,064
Rb	82	82	107	102	165	156	124	105	109	72	69
U	0	3	5	6	6	11	9	6	2	9	2
Th	12	15	8	12	21	23	10	4	4	8	11
Ce	74	77	53	28	5	5	8	6	10	83	76
La	63	68	37	48	15	18	36	36	36	58	61

a and b represent repeats of the same sample (splits of one powder).

All analyses were performed by Dr. J. G. Holland, University of Durham, England.

The key to specimen numbers can be seen in Appendix 1 iii

iii) Key to Specimen Numbers for X.R.F. Analyses

#	Specimen	Thin Section Description Name	Location of Specimen
1a 1b	5700A 5700B	Quartz microdiorite	Scranton
2	29.1	Biotite (augite?) porphyritic diorite	Upper Woodbury Creek 2.9km east of Scranton
3	29.3	Quartz diorite	Close to 29.1
4a 4b	29.6A 29.6B	Porphyritic quartz monzonite	3.3km east of Scranton Mine Mine
5	29.7	Porphyritic hornblende-biotite granodiorite	3.8km east of Scranton Mine
6a 6b	29.8A 29.8B	Porphyritic granite	4.4km east of Scranton Mine close to "Silver Spray" turnoff
7a 7b	35.2A 35.2B	Quartz diorite	Close to 29.1 at 2.8km east of Scranton Mine

Appendix 2

ATOMIC ABSORPTION

i) Principles Involved

The atomic absorption spectrophotometer (A.A.S.)works on the principle that "ground state" (i.e. neutral unbound) atoms will absorb at discrete frequencies the incident energy radiated from fixed atoms of the same element. The unknown sample is brought into solution (i.e. aspirated or dispersed) by means of a hot acetylene/air flame (temperature approximates 2000°C) which reduces the solution droplets to the atomic ground state form. In practice, a hollow cathode tube, with the cathode made of the element to be analysed, is used to supply the radiation. The amount of radiation absorbed by the ground state atoms is inversely proportional to the concentration of the element in the initial vapour i.e. the more sample element present the more absorption occurs and the resultant amount of radiation to be measured is decreased. Measurements are done by an amplifying system and a series of photomultipliers with the result being given in the form of an analogue, digital or chart recording (Levinson 1974).

All measurements are made relative to standard solutions.

The Perkin Elmer 503 A.A.S. at the University of Alberta, Geology Department was employed for all the data in this study.

ii) Analytical Procedure

a) Sulphides

Depending on the amount of powdered sample available, between 0.5 and 1 gm of pure sulphide was weighed out and dissolved in 10ml HCl and 10 ml HNO_3 (i.e. an aquae regia mixture). With an aquae regia solution, any lead in the original sulphide is precipitated as $PbNO_3$ which is soluble in aqueous solutions.

Evaporate to dryness. 10 ml of HNO_3 was then added and allowed to evaporate to dryness again. This was repeated a further two times to remove any chlorides present.

The final dried sample was then dissolved in a further 10ml of HNO_3 and the volume made up to 50ml with distilled water. Due to the high lead concentration in some of the samples it was necessary, to totally dissolve the sample, to make up the volume to 100ml, or even 200ml.

b) Granites and Individual K-Feldspars

1gm of the previously crushed sample was weighed out and then dissolved in 10ml of HF and 10ml of HNO_3 in a teflon beaker. After evaporation to dryness, this procedure was repeated a further two times.

To the dried residue, 10ml of HNO_3 was added and again evaporated to dryness. This was repeated two further times. All the granites were completely soluble in the nitric acid solution. The volumes of each sample were then made up to 50ml in a calibration flask.

A total of 12 sulphides, 6 feldspars and 7 granites were analysed.

iii) Tabulated Results a) Sulphides

		Co	Cu	Fe	Pb	Mn	Mo	Ag	V	Zn	Ni
Sphalerite	FE	25	8,005	42,411	8,358	159	0	5.3	0		0
	SA	22	212	52,525	6,235	155	0	2.7	0		0
	SG	17	60	16,529	3,625	52	0	16	0		0
	SHI	17	116	26,482	7,348	47	0	1	0		0
	SHII	12	62	20,761	11,828	32	0	12	0		0
Galena	WVC	0	384	8,816		155	0	20+?	Tr	5,360	0
	WVX	0	34	303		77	0	3.0	0	51	0
	FE	0	1,428	7,106		0	0	7.0	0	37,108	0
	SF	0		1,223		0	0	5.2	0	6,870	0
	SH	0	44	75		0	0	20+?	Tr	66	0
	SG	0	6	161		0	0	7.4	0	268	0
Pyrrhotite	WVC	66	12		10,366	417	0	5.6		1,647	12.5

All values in ppm

b) Granites and Feldspars

Granites	Co	Cu	Pb	Mn	Mo	V	Zn
5,700	95	17	34.9	831	Tr	72	112
29.1	107	29	39.6	1,043	Tr	198	99
29.3	100	14	38.4	834	Tr	106	109
29.6	144	17	36.1	213	Tr	37	38
29.7	129	11	35.6	491	0	73	71
29.8	126	10	42.4	621	Tr	73	81
35.2	104	11	33.9	753	Tr	69	107
Feldspars							
29.7F			46.9				
29.8F			61.9				
30.2F			69.4				
2700LF			41.7				
5700FI			49.4				
5700FII			55.5				
Correlation Values*							
Average Felsic Igneous	5	30	48	600	1.9	40	60

*Reported from Hawkes and Webb "Geochemistry in Mineral Exploration" 1962. Average for felsic igneous rocks.

iv) Correlation of standards with U.S.G.S. Standards

	AGVI		BCRI		SYZ		MRGI	
	A	B	A	B	A	B	A	B
Copper	59.7	54.2	18.4	17.7	6.6	5.2	135	117.2
Lead	35.1	50	17.6	35	83	105.3	12	28.3
Manganese	763	783	1,406	1,197	--	1,289	--	1,147
Molybdenum	2.3	5	1.1	5	--	--	--	--
Vanadium	125	115	399	450	46	50	--	507
Zinc	84	88.7	120	120	247	220	187	157
Cobalt*								

A Column As reported for U.S.G.S. Geological rock standards by Flanagan, 1973, Tetley and Turek 1974, Carmichael, Hampel and Jack 1968.
(NB. It should be emphasized that these values themselves vary quite largely. Wherever possible Flanagan's data was used in preference.

B Column This study

*The values for cobalt were disregarded due to contamination during the crushing procedure.

Appendix 3

SULPHUR ISOTOPES

i) Preparation of Samples

After a thorough investigation of the ore mineralogy, it was found that the sphalerite and galena appeared to be in equilibrium and therefore suitable for sulphur isotope geothermometry. Pyrite, however, often showedcrosscutting relationships and an extended paragenetic depositional sequence, indicating disequilibrium with the other sulphides.

Due to the coarse grained nature of most of the galena and sphalerite, the preparation of the mineral separates was relatively simple. In most cases it was sufficient, after careful collection, to simply check the purity under a binocular microscope. In some specimens the galena and sphalerite was too fine grained to be hand picked so that crushing had to be employed. The 100-200 mesh size was collected and the galena and sphalerite separated initially using the Franz magnetic separator. Each mineral separate was then hand picked under a binocular microscope for any minor impurities remaining.

In one specimen (5702-S) fine grained pyrite was intricately associated with the sphalerite and galena and therefore the 200-270 mesh size had to be separated.

No chemical reactions or heavy liquid separations were required for the sample preparations, thereby reducing the risk of unwanted fractionations.

The relative purity of each of the mineral separates was checked by atomic absorption spectrometry. From the following table it can be seen that the level of lead in the sphalerite reaches a maximum of 1.18% (11,828ppm) and of zinc in the galena 660ppm. Both of these values indicate that a high level of purity was attained.

Analysis of the sulphide samples for purity

Sphalerite	Lead	Iron	Zinc	Copper
SA	6,235	52,525(5.25%)		212
FE	8,358	42,411		8,005
SG	3,625	16,529		59.8
SHI	7,348	26,482		116
SHII	11,828	20,761		61.5
Galena				
WVC		8,816	222	384
WVX		303	29	34
SHI		74.8	40	44
SG		160.8	200	5.8
FE		7,106	660	1,428
SF		1,222.8	450	25

All values in ppm

ii) Collection of SO_2 Gases

Combustion and collection of SO_2 gas from the sulphides were made initially in a quartz-pyrex extraction line in the stable isotope Laboratory, Department of Geology, University of Alberta, being completed in a similar line in the Physics Department, University of Calgary.

The lines consist essentially of a quartz tube combustion chamber, pyrex tube vacuum line with two cold fingers, two thermocouple vacuum gauges to measure the vacuum, diffusion pump, roughing pump and breakseals for collecting the final SO_2 gas. Cu_2O (Fisher reagent grade), preheated to 300°C under vacuum to remove impurities, was used in the combustion as an oxygen donor. (Robinson et al 1975). The pure sulphides and Cu_2O were ground to a fine powder separately in an agate mortar and then weighed and mixed in a fixed ratio. A molar ratio of oxygen/sulphur of 4:1 (or in excess) was used (ie. approximately 24 mgm of sphalerite and 60mgm of galena to $>$160mgm of cuprous oxide).

The sulphide sample and the Cu_2O were then mixed and packed between quartz wool in a small open-ended quartz tube. The sample tube was put into the quartz chamber which was then evacuated and heated to 1000°c. During combustion the evolved SO_2, and other trace gases such as CO_2, were trapped with liquid nitrogen in the second cold finger, whilst H_2O was trapped separately by a mixture of ethanol and dry ice in the first trap. A relatively long combustion time with a high temperature was employed to minimise fractionation of the sulphur. Galena was combusted for 15 minutes and sphalerite for 25 minutes. After complete combustion the CO_2 was allowed to expand and was pumped away, while the SO_2 was frozen back into the cold finger by a mixture of n-pentane and liquid nitrogen at a temperature of about -120°C. At this temperature the vapour pressure of CO_2 is about 800 times that of SO_2 and the fractionation due to the pumping process was taken to be negligible. The SO_2 was transferred into a standard breakseal and collected.

To complete my preparation of SO_2 for a few remaining samples, the sulphur isotope line was modified at the University of Calgary so that the methods employed were the same.

iii) Mass Spectrometer Measurements

The isotope analysis were performed on the mass spectrometer in the Department of Physics, University of Calgary under the direction of Dr. H.R. Krouse.

The mass spectrometer was built around a Micromass 602 analyser, which features a magnetic valve system for alternately introducing a standard and then an unknown sample. On entering the mass spectrometer, the gas was ionised by a stable electron beam and the individual ions produced were separated according to their masses in the magnetic field. Ions of masses 64 ($S^{32}O^{16}_2$) and mass 66 ($S^{34}O^{16}_2$) were simultaneously collected. A mass of 66 can also be produced by $S^{33}O^{17}O^{16}$ and $S^{32}O^{17}O^{17}$, both of which are considered to have a negligible effect. However, with $S^{32}O^{16}O^{18}$ an oxygen correction factor has to be applied and this was taken to be the theoretical value of 1.095.

Ratios of the two ion currents were then digitally recorded and the data processed, by an online Texas Instruments 980 computer, to directly print the ratio

$$[^{34}S/^{32}S] \text{ sample} \;/\; [^{34}S/^{32}S] \text{ standard}$$

This was achieved after 9 recordings of alternate sample/standard gas introductions were made.

The formula used to calculate the final sulphur isotopic value is as follows:

$$\delta\, ^{34}S = F\left\{\frac{R_{Sample} - R_{Standard}}{R_{Standard}} - 1\right\} \times 1{,}000\ ‰$$

Where F = the oxygen correction factor 1.095 and $R = {}^{34}S/{}^{32}S$.

iv) Sample Computer Printout

```
SULPHUR MAINLINE PROGRAM NOV 1978
INSERT TWO LINE REMARK
SIMON GAL 2 STANDARD
JAN 12  2:Ø6

READY?Y
STANDARD        UNKNOWN
ØØ1.34858    ØØ1.33804
ØØ1.34831    ØØ1.33768
ØØ1.34831    ØØ1.33759
ØØ1.34818    ØØ1.33743
ØØ1.34798

RATIO   UNKNOWN STANDARD

ØØØ.99234
ØØØ.99234
ØØØ.99218
ØØØ.99211
ØØØ.99211
ØØØ.99211
ØØØ.99211

AVERAGE RATIO   ØØØ.99218   +- ØØØ.ØØØ12
```

v) Calibration with Standards

The Mayerthorpe Troilite was assigned a value of 0‰ and all the unknown samples were calibrated against this. The isotopic value of this troilite has been previously determined and found to have a true meteoritic value (personal communication Dr. A. Sasaki). Thode et al (1961) have also shown the remarkable consistency of sulphur isotopic ratios for meteorites with values falling within very narrow limits (i.e. 0‰ $\pm$ 0.2‰). Repeated runs of the Mayerthorpe Troilite, at random during the experimental procedure, gave a standard deviation of $\pm$ 0.15‰.

A galena standard was also used (NBS #200 Ivigtut PbS), which gave a value of -2.25$\pm$ 0.05‰ for 4 analyses.

After repeated runs using the Fisher reagent sphalerite standard the values appeared to be somewhat erratic. However this standard is apparently prone to give varied results (personal communication Dr. H.R. Krouse). As this standard is significantly higher than all my unknown samples, it was not used in calibration.

The system at the University of Calgary allows for a standard to be run alongside the unknown sample. After repeated runs, the standard deviation for reanalysing the gases was found to be $\pm$ 0.1‰ and for duplicate combustions to be at the worst $\pm$ 0.4‰ and usually better than 0.3‰.

Appendix 4

FLUID INCLUSIONS

i) Specimens Used

5 specimens were chosen from an original total of about 35 samples. The majority of these samples were discarded because of the small size of the inclusions and the high degree of opaqueness in white milky quartz and brown sphalerites within the study area. The specimens finally used in this study were FB (sphalerite) and FM (quartz) from Flint, PG (quartz) and PA (quartz) from the Scranton fissure system (from Pontiac) and WV#1 (quartz) from the Vigilant. The Vigilant specimens were collected from quartz vugs adjacent to the mineralised vein.

ii) Sample Preparation

Two different sample preparation methods were employed. A flat rock chip of the quartz or sphalerite sample was stuck to a glass slide using balsam cement. After cutting the specimen surface parallel to the glass slide on the thin-section machine the specimen was ground on glass plates with successively finer corundum powder. One side of the specimen was then polished on paper discs with 6 um, 3 um and then 1 um diamond paste. Care was taken to thoroughly clean the specimen between each polishing stage. The specimen was removed from the slide by soaking overnight in acetone and then remounting on the opposite side and the process repeated. The thickness of most specimens (because of the opaqueness of the quartz and sphalerite involved) was extremely low (< 0.3mm). Most of these initial samples proved unsatisfactory, because of the mounting medium involved, and therefore a second method was utilised. Here the samples were prepared similarly to a thin section being mounted in balsam on a hot plate which attained a temperature of less than 80°C and therefore would not affect the fluid inclusions. The samples were polished by a simple buffing process and then removed from the glass slide by soaking overnight in alcohol.

iii) Microthermometry

The CHAIXMECA model VT2120 combined heating-freezing stage, in the Geology Department at the University of Alberta, was used for this study. The

sample to be studied is placed on a bronze alloy sample chamber which is fitted to a microscope stage. The sample chamber is heated, using an annular resistance heating coil, the temperature being displayed on a digital voltmeter via a platinum resistance thermal-sensor mounted in the heating stage. This type of instrument was tested by Poty et al (1976) who showed that at higher temperatures in particular, a thermal gradient was produced. At 380°C a gradient of 1.0°C was produced over a distance of up to 1mm above the heating stage. It was not advisable, therefore, to use specimens thicker than a few millimetres. A small lateral gradient is also produced because of the position of the thermocouple but does not exceed 0.8°C from the centre of the optic field to its margin. The optic lenses of the microscope are cooled by a water jacket.

Careful calibration of the instrument is necessary using organic liquid standards generally below 0°C and inorganic standards above 0°C.

A maximum heating rate of 0.25°C per minute, when within 10°C of the target temperature, has been suggested by the Chaixmeca Corporation.

Cooling is effected by passing a stream of nitrogen gas through a copper coil within a liquid nitrogen bath and then to the freezing stage below the sample chamber.

The resolution of the optics was found to be much reduced at both high temperatures and also when condensation occurs on the specimen at low temperatures. As a result, during the heating experiments a glass window shield was used to protect the lens and during freezing experiments a plastic sleeve was placed over the lens to try and reduce the condensation.

iv) Crushing Experiments

The gases within the fluid inclusions were tested, using a simple crushing stage similar to that described by Roedder (1970b). The device consists essentially of two metal plates hinged at one margin and joined by a vertical screw at the other. The sample to be studied is placed between the metal plates, on glass slides, viewed through a circular hole in the upper plate onto which a small lens is attached. By increasing the pressure exerted on the sample by rotating the screw,

any gases are released into the surrounding medium. To analyse the gases, the sample is embedded in various media depending on the gas expected to be produced (eg. glycerin as an all-purpose medium, CO_2 saturated kerosene for organic vapours, barium hydroxide for CO_2). The rate at which the gas dissolves in the medium is an indication of the type of gas present. Care must be taken when mounting the specimen between glass slides so that it is fully embedded within the medium.

v) Volumetric Salinity Determinations

For inclusions in which halite (and sylvite?) were present as daughter crystals the salinities were calculated by measuring the volumetric proportions of the various phases within the fluid inclusions. At low volume percentages, however, it is often difficult to determine exactly the precise volume of each of the phases because of:-

i) difficulty in focusing of the vapour phase.
ii) correction for a spherical (vapour) or cubic (halite) form.

As described by Kelly and Turneauve (1970) the salinity of a solution can be determined from the following equation:

Total NaCl weight equivalent = weight of solid + gas + fluid phases.

The gas phase is considered to be negligible in the calculation and therefore the volume of liquid = 100% - (volume of salt + volume of vapour).

The weight of the salt (solid) = 2.164 (for halite) x volume of salt where the density of halite is 2.164gm/cc.
The weight of the fluid (brine) = 0.317 (volume of liquid x 1.199) where the brines enclosing only halite crystals were assumed to contain 0.317gm NaCl per cc and have a density of 1.199gm/cc at room temperature.
If sylvite and halite are present the brines have a density of 1.234gms per cc and contain 0.126gm KCl and 0.258 gm NaCl.

Work by Clynne and Potter (1973) has shown that the weight % NaCl equivalents calculated from the above equation are generally within $\pm$ 5% of the actual concentration within more complex brines.